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SUBMARINE GEOLOGY OFF SAN DIEGO, CALIFORNIA
K. O. Emery, W. S. Butcher, H. R. Gould, and F. P. Shepard 511

PHASE RELATIONS OF THE ALKALI FELDSPARS. II. THE STABLE AND PSEUDO-STABLE PHASE RELATIONS IN THE ALKALI FELDSPAR SYSTEM . . Fritz Laves 549

PLEISTOCENE EVOLUTION OF THE OHIO AND WABASH VALLEYS . William J. Wayne 575

THE 10 PER CENT $CaAl_2Si_2O_8$ PLANE IN THE SYSTEM $CaSiO_3$-$Ca_3Al_2SiO_7$-$NaAlSiO_4$-$CaAl_2Si_2O_8$ . . . . . Hatten S. Yoder, Jr. 586

REVIEWS 594

INDEX 601

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November 1952 Vol. 60, No. 6

THE JOURNAL OF GEOLOGY

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BEGINNING with the January issue, the *Journal of Geology* will be edited by Leland Horberg. Dr. Horberg has been a contributing editor these last five years and has given generously of his time in the service of the *Journal*. My thanks to readers, contributors, and co-workers of the *Journal* who have borne with me during the last five years and my best wishes to the incoming editor.

F. J. PETTIJOHN

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SMITH, A. V. (1926) Discovery of fish remains in Ordovician rocks of the Black Hills (abstr.): Geol. Soc. America Bull. 52, p. 27.

——— and DUNSANY, A. J. (1929*a*) Early Ordovician faunas of South Dakota: U.S. Geol. Survey Bull. 444, pp. 1–76.

——— — ——— (1929*b*) Revision of Ordovician stratigraphy of the Black Hills: Jour. Geology, vol. 37, pp. 680–695.

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THE JOURNAL OF GEOLOGY

November 1952

SUBMARINE GEOLOGY OFF SAN DIEGO, CALIFORNIA[1]

K. O. EMERY,[2] W. S. BUTCHER,[3] H. R. GOULD,[2] AND F. P. SHEPARD[3]

ABSTRACT

The coast in the vicinity of San Diego is bordered by a shelf only 5–10 miles wide. Near the southern limit of the area, several small islands rise from the outer part of the shelf. These islands have a submarine extension to the north as a low ridge that is cut by a transverse canyon. North of the canyon the ridge forms an isolated, flattened bank that continues as a narrow ridge northward into deep water. Separating the bank and deep ridge from the shelf is a submarine valley having a "drainage" divide in the southern end. North of the divide the valley floor slopes northwestward into San Diego Trough. South of the divide the valley slopes southward, meeting the transverse canyon at a point east of the ridge. An escarpment just west of the islands and the bank forms one side of a deep graben-like depression known as San Diego Trough. A delta-like feature at the mouth of the canyon extends from the foot of the escarpment partway across the floor of San Diego Trough. The relationship of the transverse and longitudinal valleys and of the flat-topped "delta" may provide new information on the origin of submarine canyons.

On the inner part of the shelf south of Point Loma is an area of outcropping Cretaceous shale and sandstone only partially covered with unconsolidated sediment. The bank top has another large tract of bedrock that consists of sandstone and shale of late Pliocene to late Pleistocene age, overlain in part by nodules of phosphorite which are Pleistocene to Recent in age. Rock cropping out on the shelf around the islands is lithologically similar to the Miocene rocks of the islands.

Gravels are abundant in the various areas of bedrock outcrops and also are present in some areas where no bedrock is known. One tract of boulders and cobbles with only interstitial sand occurs a mile off the mouth of the Tía Juana River; similar gravels, but with much more sand, form an elongate area south of Point Loma. Most of the gravels are covered by delicate growths, showing that they are not being moved at the present time. Possibly they are residual deposits formed during times of glacially lowered sea-level.

The finer sediments that partially blanket the sea floor have a distribution that is much more complex than the usual concept of marginal marine sediments. The chief factors that control the distribution are: (1) earlier deposition of coarse sediments at depths now below the effective reach of wave action, (2) present-day currents that by-pass fine detrital sediments and allow accumulation only of coarse calcareous organic sediments, and (3) present-day deposition of fine sands in areas adjoining beaches and of silts in deeper water. The mineralogy of the inner shelf and bank sediments is similar to that of the Pliocene-Pleistocene rocks that crop out on land and on the sea floor; hence they are considered as having been derived from Pliocene-Pleistocene rocks or from the same sources that supplied the Pliocene-Pleistocene sediments. Clastic sediments in the vicinity of Los Coronados Islands appear to be locally derived from the Miocene rocks of the islands.

INTRODUCTION

Acoustic studies conducted by the University of California Division of War Research at the United States Navy Radio and Sound Laboratory in San Diego, California, required the collection of numerous samples from the sea floor. Most of these were obtained off San Diego during the period April, 1942, to February, 1945. Because of their dense spacing, the samples yield a picture of sediment distribution that has been

[1] Contribution of Scripps Institution of Oceanography, new series, no. 588. Contribution of Allan Hancock Foundation no. 85. Manuscript received December 18, 1951.

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[3] Scripps Institution of Oceanography, La Jolla, California.

equaled at few other places in the world. The main suite has been supplemented by prewar and postwar samples taken in portions of the area where acoustic studies were not made.

Altogether, 1,656 samples are available. These include 132 large dredge hauls; the remainder are smaller samples taken with snappers or scoopfish (Emery and Gould, 1948). In addition, there are 184 usable photographs of the bottom, some of which have been described earlier by Shepard and Emery (1946).

Each of the authors participated in both collection and study of the samples but made his main contribution to one phase of the work. Shepard planned and conducted the field operations and is responsible for most of the physiographic interpretation, Butcher studied the rocks, Gould investigated the minerals in the sediments, and Emery made other sediment analyses and brought the various results together into this report. Because of the very large number of samples and the comprehensive scope of the work, this was necessarily a co-operative project. The work was aided by many other persons, chief among whom are Dr. Orville L. Bandy, Mr. R. D. Briggs, Mrs. Jane Dinkens, Mrs. Anne King, and Miss Frances L. Parker. Funds for the field work and some of the laboratory studies were provided by the University of California Division of War Research. Additional laboratory studies were made possible by fellowships granted by the Allan Hancock Foundation, by the Scripps Institution of Oceanography, and by Shell Oil Company.

Topography

Sources of data

Surveys by the United States Coast and Geodetic Survey during 1934 and 1935 gave a good coverage of the nearshore area from Point Loma to the Mexican border and provided additional information along portions of the outer shelf and the slope down into San Diego Trough. Scattered soundings south of the Mexican border were available on Hydrographic Office charts. However, the needs for the acoustic work were so exacting that additional soundings were required for all except the inner and the northern portions of the area. These were obtained on the "E. W. Scripps" with an echo-sounder of the old visual flashing-light type. Since the war the surveys have been greatly augmented, and details have been added in critical areas. This recent work has been a co-operative undertaking between the United States Navy Electronics Laboratory and Scripps Institution of Oceanography. The ships involved have been the "E. W. Scripps," the "Horizon," and the Navy ships "EPCE(R) 855" and EPCE(R) 857," all of which are equipped with recording echo-sounders (pl. 1, *C*).

Positions for the surveys were obtained by horizontal sextant angles be-

PLATE 1

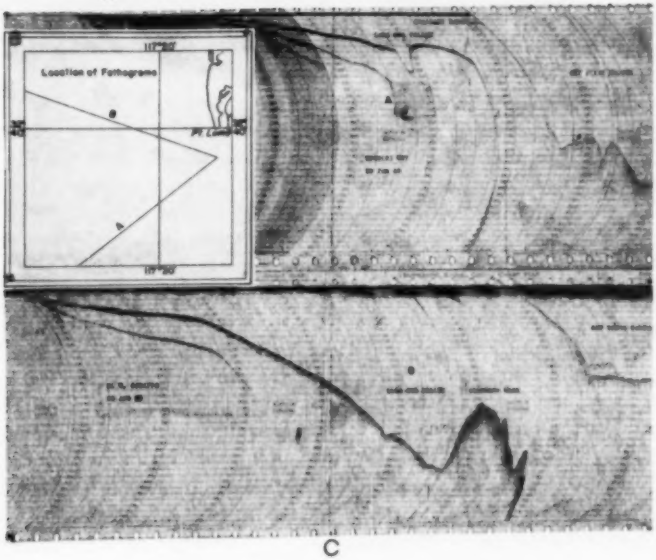
A, Underwater photograph, showing truncated edges of bedrock covered with attached organisms, mostly brachiopods. Bottom edge about 4 feet long. Lat. 32°32′9, long. 117°20′2, 62 fathoms. University of California Division of War Research.

B, Underwater photograph of rounded cobbles covered with attached organisms and partly buried in foraminiferal sand. Bottom edge about 4 feet long. Lat. 32°32′5, long. 117°19′9, 67 fathoms. University of California Division of War Research.

C, Typical fathograms off San Diego. Note the scale change on the seaward side of Coronado Bank: 400 fathoms should be added to the indicated depths in San Diego Trough.

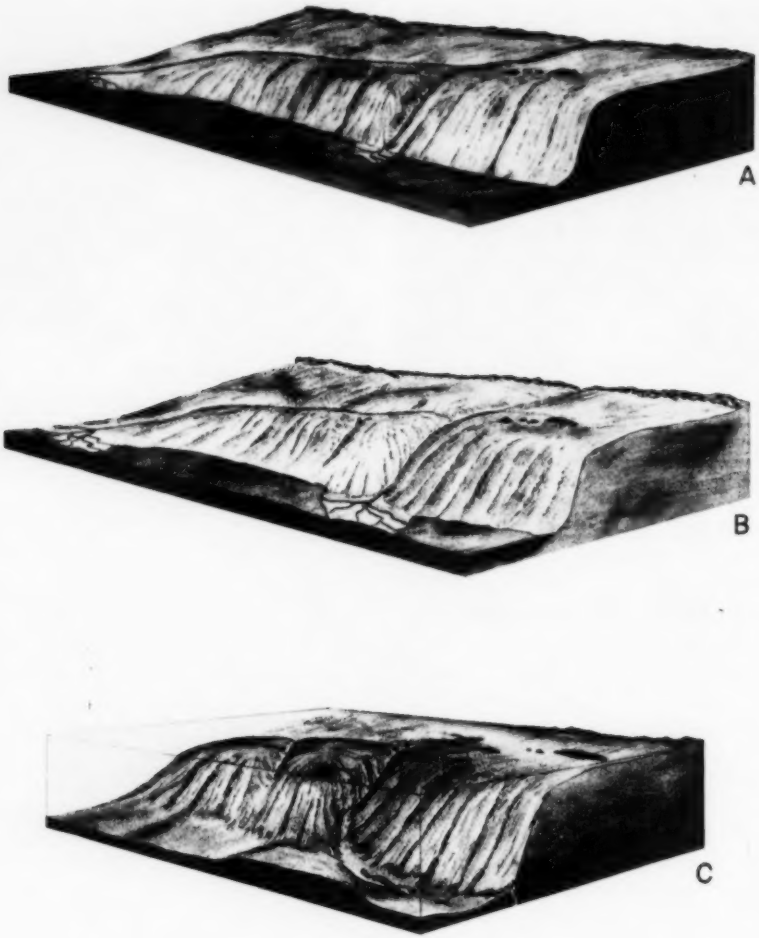
A

B



C

Underwater photographs and typical fathograms



Hypothetical development of Coronado Canyon

tween accurately located points on shore. In some of the outer portions of the area, where angles are small and visibility poor, parts of the sounding lines could be located only by dead reckoning. The surveys are now so complete that there is little difficulty in navigating by echo-sounding, even in dense fog.

Contours based on all the soundings were drawn at 10-fathom intervals from shore to a depth of 100 fathoms, and at 50-fathom intervals below that depth (fig. 1). These contours show that the submarine area of 760 square statute miles can be divided into several sections: shelf (45 per cent), Loma Sea Valley (10 per cent), Coronado Canyon (1 per cent), Coronado Bank (5 per cent), escarpment (11 per cent), and San Diego Trough floor (28 per cent).

SHELF

A somewhat complex shelf constitutes nearly half the area. South of latitude 32°30′ it approximates the ideal continental shelf of the world, with a gradually increasing slope from shore to the break in slope at a depth of about 70 fathoms, 10 miles at sea. The average declivity from 0 to 30 fathoms is only 0°10′10″ (18 feet per mile), whereas from 30 to 70 fathoms it is four times as steep. The smoothness of its surface is broken by Los Coronados Islands, a series of small tilted fault blocks that rise along the outer portion of the shelf, and their northward extension as a low submarine ridge.

North of latitude 32°30′ the shelf is quite different. Ranging between 5 and 10 miles in width, it is steeper and has only a slight break in slope at about 60 fathoms, somewhat shallower than farther south. Beyond that depth the bottom has an average declivity of 2°30′ (230 feet per mile) to the axis of Loma Sea Valley. Just south of Point Loma the shelf is interrupted by a relatively shallow, flattened area that is bounded on the west by a 10-fathom escarpment, the latter extending northward along Point Loma a mile out from land.

CORONADO BANK

Lying between Loma Sea Valley and San Diego Trough is a fairly flat area having depths similar to those of the outer part of the shelf. Because of its shallow depth, the area in a sense belongs to the shelf, but, because of its separation from the true shelf by the deep Loma Sea Valley and by Coronado Canyon, it is termed "Coronado Bank." Owing to its relative geomorphic isolation, this area of 37 square miles has sediment which differs markedly from that of the true shelf.

The top of the bank has two high areas, separated by a saddle. The southern high has a flattened top at a depth of about 55 fathoms and is bordered by gentle slopes except to the south, where the top is sharply trenched by Coronado Canyon. The flattened top of the northern area has an average depth of about 75 fathoms. The western edge of both parts of the bank has a break in slope varying from 80 to 90 fathoms, and the eastern edge is somewhat shoaler in most places. In the saddle between the highs, depths

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PLATE 2

A, Exposure of slope and beginning of subaerial erosion.
B, Piracy of head of Loma Sea Valley by steeper Coronado Canyon. Development of delta.
C, Rise of sea-level and dissection of delta by slumping and turbidity currents.

range down to 100 fathoms. Probably the flattening of the bank top was produced during times of lower sea-level of the Pleistocene epoch, but subsequently the surfaces must have undergone some warping.

LOMA SEA VALLEY

Loma Sea Valley is fairly straight and parallel to the general structural trend of the region. It has an asymmetrical cross section with average slopes of 2°30′ on the east and 15° on the west between the 100- and 200-fathom contours. The shallow south end of the valley is more symmetrical (pl. 1, *C*). The much faulted Middle and South Coronados Islands are directly in line with an extension of the axis of the valley. Consequently, the valley may have been eroded along a fault line, or it may be of structural origin. Landward of the shoalest portion of Coronado Bank the valley has a divide from which its axis slopes away in both directions, running directly into San Diego Trough to the north and into Coronado Canyon to the south. An erosional origin or erosional modification of the valley is suggested by its more or less continuous outward slope and by the finding of a large quantity of cobbles in a dredging along its axis.

CORONADO CANYON

One of the most interesting topographic features of the area is Coronado Canyon. It is transverse to the general structure of the region and extends landward to within 7 miles of the coast near the Tia Juana River mouth. The landward portion of the valley reaches depths as shoal as 37 fathoms, where its gently sloping walls are apparently free of rock. Farther west, the canyon has steep rocky walls and a winding course. Several tributaries join it east of the ridge formed by Coronado Bank and the northward continuation of Los Coronados Islands. In its "water gap" through the ridge the canyon has a straight course and attains its maximum depth of 340 fathoms below surroundings with side slopes as steep as 33°. Photographs (Shepard and Emery, 1946, fig. 29) show small projecting or overhanging rock layers along the upper slopes and truncated edges of tilted beds (pl. 1, *A*). West of the gap, where the floor of the canyon has a depth of about 500 fathoms, a change in character takes place (pl. 2). The axis turns south, and a trough with blunt-headed tributaries appears. In places along this outer valley the channel divides and rejoins, leaving hills in the center. Locally, the channel is flanked by low ridges similar to those that have been described by Menard and Ludwick (1951) along the outer parts of other canyons. This peculiar troughlike valley probably crosses the southern end of San Diego Trough, and it may continue into San Clemente Basin beyond. West of the point where Coronado Canyon changes character, there is a delta-like feature, into the east side of which the outer part of the valley appears to have been cut. The top of the "delta" has a relatively flat surface at about 500 fathoms that forms an arc seaward of the mouth of the main canyon.

The usual difficult problem of origin of submarine canyons exists in the San Diego area. There is, however, some important new evidence which comes from the detailed soundings along this canyon system. The relation of Loma Sea Valley to Coronado Canyon is a feature of special significance. If this portion of the sea floor had once been subject to subaerial erosion due to greater elevation in the past, it may have been possible for a stream working headward in a saddle in

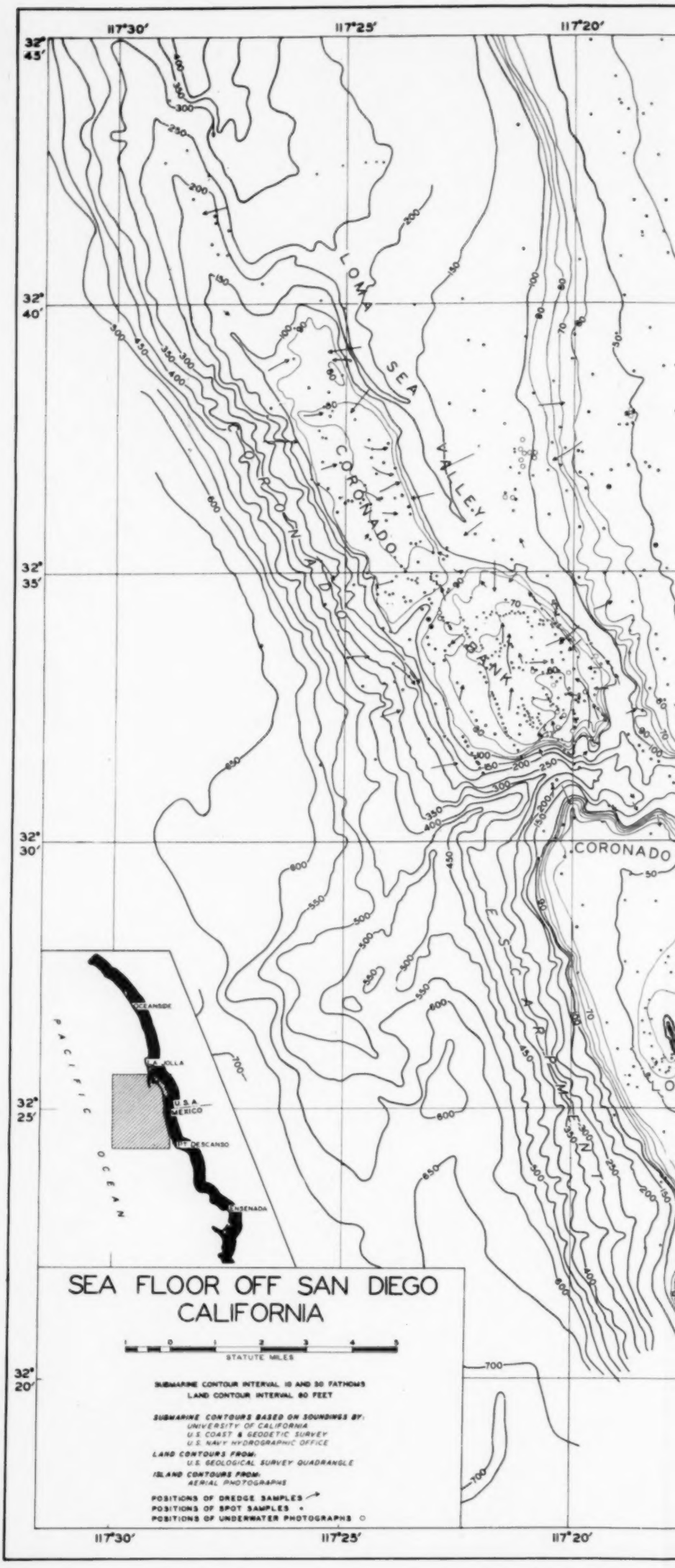
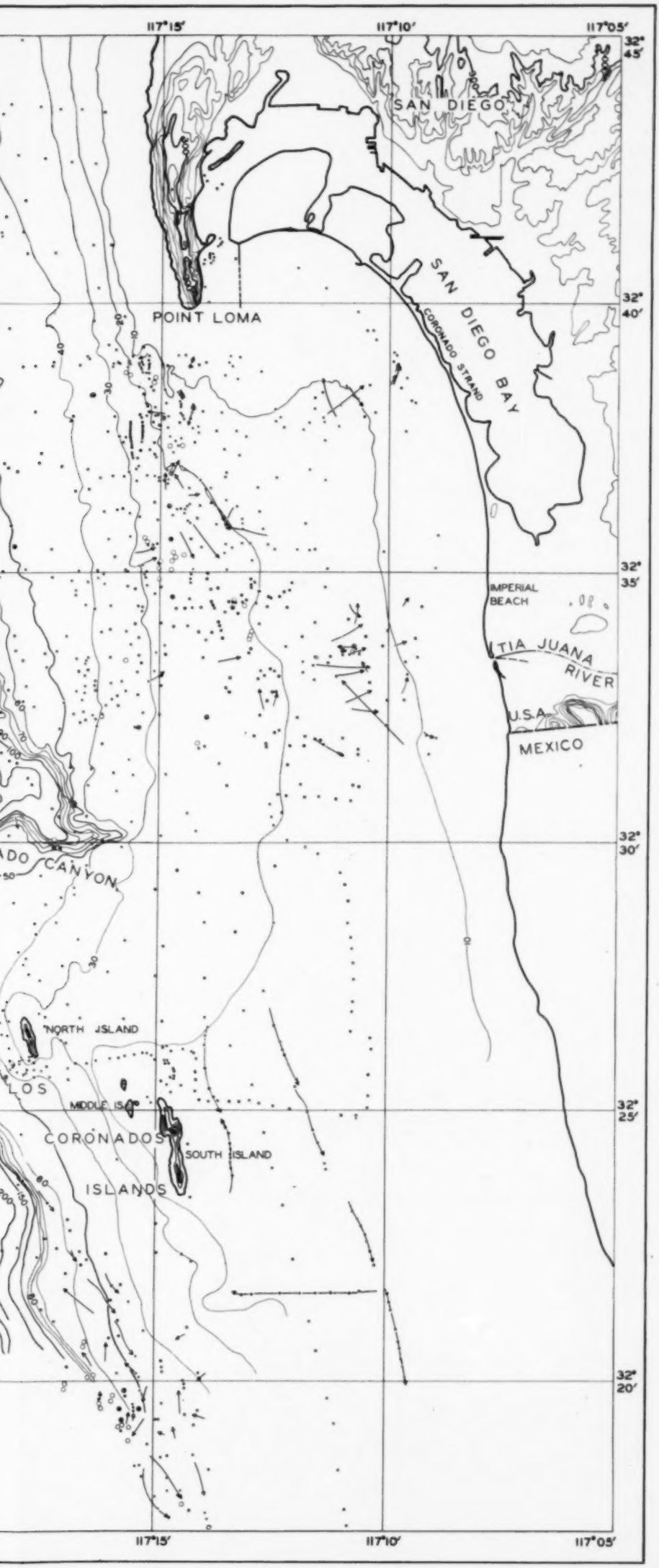


FIG. 1.—Bottom contour chart of area



f area off San Diego, California

the outer ridge to capture another stream flowing down what is now Loma Sea Valley (pl. 2). Subsequently, the full drainage from large valleys on shore, such as the Tia Juana and San Diego rivers, could have flowed across the ridge and cut the deep "water gap" now found submerged. The delta-like surface at the mouth of the canyon could have been formed where the stream entered the sea at this level. It seems unlikely that turbidity or other types of currents on the sea floor could have produced these typical stream-erosion features; there would have been no obvious source of sediment for the turbidity currents on the outside of the ridge prior to the cutting of the gap, nor would these currents account for the horizontal top of the "delta."

The peculiar outer valley with its depressions and low bordering ridges may have been initiated by a partial collapse of the delta after submergence. Turbidity currents coming out of the rocky gorge may have flowed down a landslide depression in the delta front, producing more of a valley appearance than would have existed previously. The presence of the low ridges along the outer valley is particularly significant, because similar features exist at the sides of channels on the fore-set slopes of deltas built into Swiss lakes (Kuenen, 1950, fig. 227).

ESCARPMENT

Beyond the break at the outer edge of the shelf and the bank is a slope leading down into San Diego Trough. This is comparable to the continental slope in its relationship to the adjacent shelf but cannot be termed a true continental slope because it does not mark the boundary between the continent and the deep sea floor. The real continental slope is about 140 miles farther seaward. Between the depths of 65 fathoms and about 500 fathoms the escarpment has an average width of 2.5 miles and a declivity of 13°. On both sides of the delta the escarpment meets the floor of the San Diego Trough at about 650 fathoms.

Closely spaced echo-sounding profiles along the slope show that a series of small and somewhat discontinuous valleys cut the escarpment. The largest, near the saddle in Coronado Bank, reaches a maximum depth below surroundings of about 100 fathoms in the middle portion of the slope, but rapidly decreases in relief both up- and downslope. The other valleys are smaller, mostly less than 40 fathoms below their surroundings. It seems likely that most of these valleys extend down to the base of the escarpment (fig. 2), but they possess few tributaries and have little in common with the larger submarine canyons. Two possible origins may be considered. One is that they are the result of landslides and turbidity currents on the escarpment, and the other is that they are drowned river valleys. Since there is no evidence of rock walls, slides in the overburden may be adequate. On the other hand, if the suggested subaerial origin of Coronado Canyon is correct, other valleys should have been cut into the outer slope during emergence. Such valleys may have been maintained or reopened to some extent by landslides and turbidity currents after their submergence.

SAN DIEGO TROUGH

The area of figure 2 includes only a small portion of the San Diego Trough. The maximum depth shown is about 720 fathoms. The floor, which extends beyond the chart, is so wide and flat that it would appear to be the top of a thick fill of sediment. Seismic measurements made by Dr. R. W. Raitt (1949) show a fill of

7,000–9,000 feet above basement rocks in the part of the trough directly off San Diego, supporting this view. A series of small deeply submerged hills, located a few miles southwest of the corner of figure 2, separate the south end of the trough from the adjacent San Clemente Basin. These hills are so close together that the sediment between them cannot be very thick. Thus they probably mark the position of the sill of a former San Diego Basin that has become filled to overflowing with sediment.

Waves and Currents

Wave turbulence that places sediment in suspension and currents that transport the suspended material are important in controlling the type and distribution of bottom materials. A very general seasonal picture of the water characteristics can be obtained from a summary of oceanographic data collected over the San Diego Trough between 1942 and 1945 and of meteorological data from Scripps Institution, located a few miles north of the area studied (fig. 3).

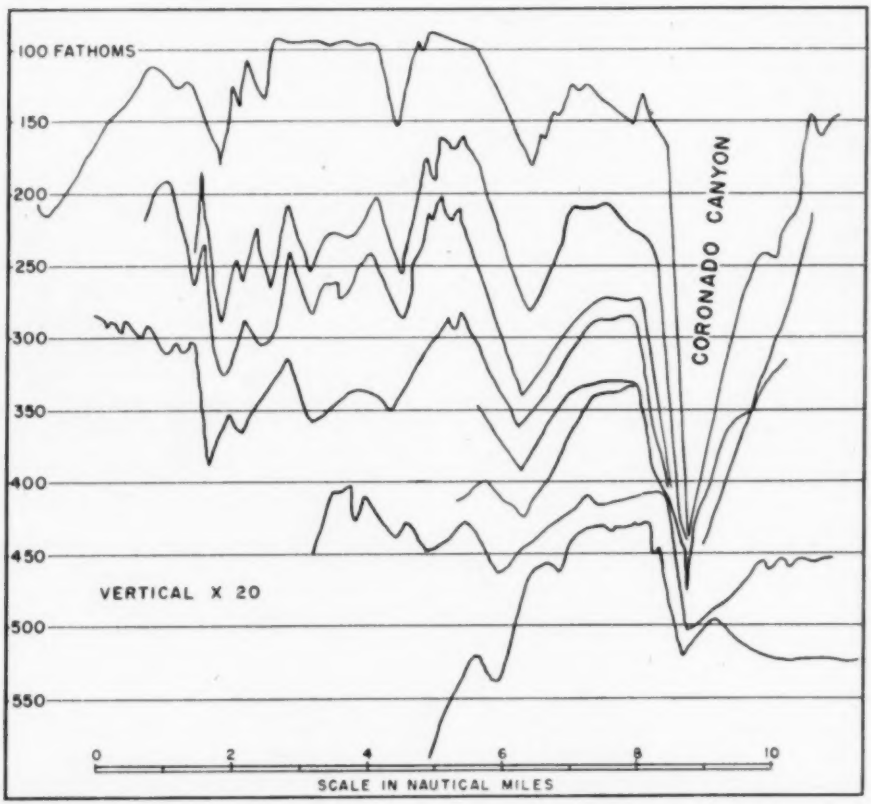


Fig. 2.—Sounding sections along face of Coronado Escarpment, showing presence of small valleys as well as of large Coronado Canyon.

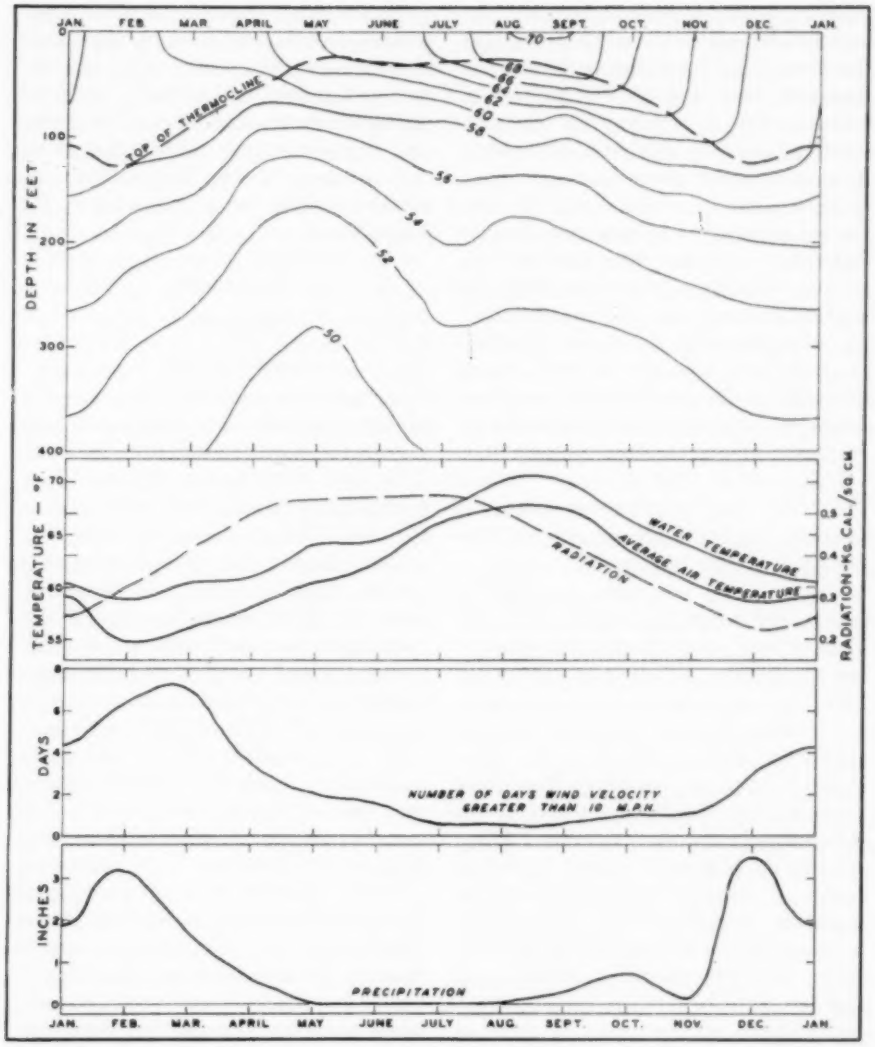


FIG. 3.—Annual variation of temperature of water and air off San Diego. Wind velocity and precipitation are also given.

Wave turbulence is a function of wind velocity, which is highest between December and April. During these months the surface water becomes well mixed to depths of more than 100 feet and cooled to about 59° F. During the summer months, when the wind is light, heating by solar radiation raises the temperature of the surface water to about 70° F., and the mixed layer is usually less than 30 feet thick; thus less disturbance of the bottom should result in summer than in winter. The months of greatest turbulence coincide with the period of annual precipitation. Streams are then intermittently active and carry to the ocean sediment which is soon redistributed by currents. Internal waves in this area have velocities of at least 0.7 knot (Ufford, 1947), but their effect on sediments is unknown because no measurements near the bottom have been made.

Currents are the main transporting agents, and, where they are of sufficiently high velocity, they produce enough turbulence to place sediment in suspension. Tidal currents usually attain their greatest velocity at narrow entrances to bays, and, accordingly, coarse sediment occurs in the entrance to San Diego Bay. Tidal currents may also be important at the outer edge of the shelf (Fleming, 1938). In addition, a general oceanic current located well offshore results from regional variations in temperature and salinity. Surveys by Sverdrup and Fleming (1941) prove that the prevailing direction of flow is southerly, and the usual quarter- to half-knot southerly set of ships on east-west courses off Point Loma shows that a fringe of this current overlaps the outer edge of the shelf. The concentration near the edge of the shelf of both the tidal and the oceanic currents is probably the chief cause of the coarsening of sediment near Los Coronados Islands and on Coronado Bank. Other currents near the shore are chiefly due to the oblique approach of waves, and these are important in controlling the distribution of sediments supplied by streams. Shepard (1950) has shown that this longshore current is predominantly southeasterly near the entrance of San Diego Bay and northerly near the mouth of the Tia Juana River.

LITHOLOGY

MAINLAND

SEDIMENTARY ROCKS

Sedimentary rocks on land range in age from Cretaceous to Pleistocene. Late Cretaceous (Chico) rocks crop out at La Jolla and Point Loma (Hanna, 1926; Hertlein and Grant, 1944) and south of the international boundary near Ensenada (Beal, 1948). In each area they consist chiefly of yellow to brown sandstone with thin beds of gray shale. Some conglomerates, rare limestones, and carbonaceous layers are included. The sandstones are characterized by an abundance of quartz and feldspar, with heavy minerals dominated by biotite and green hornblende (table 1). Glaucophane is very rare or absent. Neither micro- nor macrofossils are common in the formation as a whole.

Eocene beds resting unconformably on Cretaceous have been divided by Hanna (1926) into four members. Lowest of these is the Delmar sand, consisting of green, gray, purple, and red sands and sandy shales, with some thin beds composed almost entirely of oyster shells. The fossils are characteristic of brackish water. Next is the Torrey sand, massive, white, and probably nonmarine. The overlying Rose Canyon shale consists of light-gray mudstone, gray shale, fine sandstones, and lenses of conglomerate

near the top. The sandstones range from light to dark brown and are commonly cemented by calcite. Pebbles and cobbles of the conglomerate are mostly well-rounded metavolcanic rocks. The topmost member, the Poway conglomerate, is similar to the conglomerate of the upper Rose Canyon. It includes some lenses of cross-bedded yellow and brown sands and gray shales. Much of this formation appears to be of continental origin. Lithologically, the Eocene rocks resemble those of the Cretaceous, but generally they may be distinguished by their poorer consolidation. Unlike those in the Cretaceous sands, the heavy minerals of the Eocene sands are usually high in epidote and sphene and low in biotite and green hornblende (table 1).

Oligocene rocks appear to be absent and Miocene rocks, other than the Comondu formation, are unknown on the mainland nearer than 30 miles north of San Diego and several hundred miles south of it. The Comondu formation, predominantly of Miocene age, covers a wide area south of the international boundary directly inshore from Los Coronados Islands (Beal, 1948). It is highly varied, consisting of rhyolitic, andesitic, and basaltic lava flows, agglomerates, and tuffs with some conglomerates, sandstones, clays, and limestones. North of San Diego, Miocene rocks are characterized by diatomaceous cherty shales, mudstones, and sandstones. The San Onofre breccia consists of breccias and conglomerates of amphibole schist and saussurite gabbro (Woodford, 1925). Glaucophane is the most outstanding heavy mineral of the schists.

Pliocene rocks cover a large area on land, being known as the San Diego formation in the United States (Hertlein and Grant, 1944) and as the Salada formation just across the border in Mexico (Beal, 1948). The Pliocene generally consists of poorly cemented, fine to coarse-grained, gray to brown sandstones with some shales, shell beds, marls, and conglomerates. The heavy minerals appear to be similar to those of the Eocene rocks, except that actinolite and hornblende are more abundant (table 1). Atop the widespread Linda Vista terrace is a thin capping of red and brown conglomerate and sand, known as the "Sweitzer formation" (Hertlein and Grant, 1944).

Gravels, sands, muds, and shell beds of Pleistocene age occur throughout the land portion of the area shown in figure 1. Their poor cementation permits them to be broken down easily and redeposited with modern sediments.

METAMORPHIC ROCKS

Metamorphic rocks occur within the basement complex east of San Diego and in the central area of Baja California. These rocks are considered Triassic and Jurassic in age (Hanna, 1926). The nearest are the Black Mountain series (renamed the "Santiago Peak series" by Larsen, 1948) of metamorphosed trachytic and andesitic flows, agglomerates, tuffs, with conglomerates, sandstones, and shales. Coarser schists and gneisses, possibly of Paleozoic age, occur in the high mountains farther east of San Diego. No outcrops of the Jurassic Franciscan glaucophane schists are known in the region. Irregularly altered sedimentary and igneous rocks, grouped as the San Fernando formation of Cretaceous age, crop out 100 miles southeast of San Diego.

IGNEOUS ROCKS

A large composite batholith of middle and late Cretaceous age (Woodford and Harriss, 1938) forms the core of the peninsula of Baja California and the high

TABLE 1

MINERAL COMPOSITION OF ROCKS IN THE SAN DIEGO AREA

MINERAL SPECIES	MAINLAND													CORONADO ISLANDS		OFFSHORE			
	Pre-Tertiary Volcanics, Agglomerates, and Tonolite Batholith, East of San Diego, Calif.*	Cretaceous—La Jolla, Calif. (Av. of 4 Samples)	Cretaceous—Ft. Banda, Lower California, Mexico	Eocene—Del Mar Sand, La Jolla, Calif.	Eocene—Torrey Sand, La Jolla, Calif. (Av. of 2 Samples)	Eocene—Rose Canyon Shale, La Jolla, Calif. (Av. of 2 Samples)	Eocene—Rose Canyon Shale, near Mercy Hospital, San Diego, Calif.*	Pliocene—San Diego Fm., Pacific Beach, Calif.	Pliocene—San Diego Fm., near Mercy Hospital, San Diego, Calif.*	Pliocene—San Diego Fm., India and Upas Sts., San Diego, Calif.*	Pliocene—San Diego Fm., near International Boundary*	Pliocene Pleistocene? Sweitzer Fm., Balboa Park, San Diego, Calif.*	Beaches—Southern and Lower California, Because Its Organisms (Av. of 6 Samples)†	Miocene—South Coronado Island (Av. of 3 Samples)	Miocene—North Coronado Island	Cretaceous—Shale, South of Point Loma	Eocene?—Sandstone, Coronado Island Area	Pliocene?—Sandstone, Coronado Bank (Av. of 4 Samples)	Pliocene?—Shale, Coronado Bank (Av. of 3 Samples)
Light minerals:																			
Quartz	X‡						54.9%		41.8%	50.5%	52.3%	50.2%	46.2%						
Plagioclase	X‡						25.0		24.9	12.1	16.8	11.1	31.5						
Orthoclase	X‡						19.0		29.9	25.3	19.8	35.2	22.3						
Calcite	X‡																		
Sericite	X‡																		
Rock fragments												3.0							
Heavy minerals: †																			
Green hornblende	X	27%	p		1%	10%	1.1%	35%	3.4%	12.1%	11.1%	0.5%	74.7%	1%	p	1%		13%	1%
Brown hornblende		p‡			p				RR	R	R						p	4	2
Actinolite		p	1%		p			9	A	MA	MA	A	1.3	1					
Magnetite	X						VR‡		R	R	R	A	2.0						
Ilmenite							MA‡	7	MA	MA	MA	A	1.3						
Opaque		4	35	27%	14	22								34	38%	35	23%	14	14
Barite													0.4	p					
Hypersthene		p				p							2.3					1	
Biotite	X	57	15	3	p	11		18	RR	RR	A	A	2.8	6	17	3	6	27	36

* Reported by G. A. Macdonald in Hertlein and Grant (1944).

† Reported by J. F. Gale (Master's thesis, University of Southern California).

‡ Description of symbols: "X" indicates present, but relative abundance has not been determined; "p" indicates present in amounts less than 1 per cent; "A" = abundant, "MA" = moderately abundant "RR" = rather rare, "R" = rare, "VR" = very rare.

TABLE 1—*Continued*

Mineral Species	Mainland												Coronado Islands		Offshore			
	Pre-Tertiary Volcanics, Agglomerates, & Tonalite Inliboith, East of San Diego, Calif.*	Cretaceous—La Jolla, Calif. (Av. of 4 Samples)	Cretaceous—Pt. Banda, Lower California, Mexico	Eocene—Del Mar Sand, La Jolla, Calif.	Eocene—Torrey Sand, La Jolla, Calif. (Av. of 2 Samples)	Eocene—Rose Canyon Shale, La Jolla, Calif. (Av. of 2 Samples)	Eocene—Rose Canyon Shale, near Mercy Hospital, San Diego, Calif.*	Pliocene—San Diego Fm., Pacific Beach, Calif.	Pliocene—San Diego Fm., near Mercy Hospital, San Diego, Calif.*	Pliocene—San Diego Fm. India and Ulas St., San Diego, Calif.*	Pliocene—San Diego Fm., near International Boundary.*	Pliocene Pleistocene? Sweitzer Fm., Balboa Park, San Diego, Calif.*	Beaches—Southern and Lower California, Descanso to Oceanside (Av. of 6 Samples)†	Miocene—South Coronado Island (Av. of 3 Samples)	Miocene—North Coronado Island	Cretaceous—Shale, South of Point Loma	Eocene?—Sandstone, Coronado Island Area	Pliocene?—Sandstone, Coronado Bank (Av. of 4 Samples)	Pliocene?—Shale, Coronado Bank (Av. of 3 Samples)
Heavy minerals—(Continued):																		
Epidote	X	2%	15%	31%	36%	9%	A‡	7%	MA	RR	RR	A	0 6%	12%	12%	p	p	2%	1%
Zoisite		p			p	1	A		MA	RR	RR	RR	0 7	p					
Garnet		2	1	3	3	1	RR‡	6	MA	R	R	RR		3	1		5%	1	p
Sphene	X	3	8	6	13	11	RR	8	MA	MA	MA	R	1 3	8	1	p	11	3	p
Apatite	X	p	p			1								p			2	p	p
Leucoxene													0 4						
Chlorite	X	2	1	15		20							0 2	2	20	1%	2	20	19
Basalt. hb.													0 5						
Zircon	X	1	3	1	4	3	R‡	1	MA	MA	MA	MA			p	p	4	2	p
Glaucophane						p			RR	R		MA	0 3	p				p	p
Olivine													1 3						
Enstatite													1 1						
Hematite													0 2						
Limonite	X																		
Alteration		1	19	13	28	10		8					4 8	31	9	58	45	11	24
Augite	X											MA	0 8						
Diopside	X												0 2						
Rutile		p			p					R	R	RR	0 1						
Andalusite		p					R												
Pyrite	X												0 3						
Uralite	X																		
Piedmontite							RR												
Tourmaline				1		p		p				R	0 8	p					
Topaz													0 6						
Collophane													1 0						
Clinozoisite																			p
Unknown		p			p														

country east of San Diego. Individual intrusive units range in composition from gabbro to granite with tonolite, gabbro, and granodiorite being especially widespread. A typical tonolite consists of andesine, quartz, biotite, and hornblende, with sphene, zircon, magnetite, and apatite as accessory minerals.

Aplite and pegmatite dikes cut the batholith and associated metamorphic rocks. These contain various semiprecious gem minerals, such as tourmaline, spodumene, and beryl. Postbatholithic rocks are limited to a few mafic dikes and andesite plugs.

Cenozoic volcanic rocks have been described with the Comondu formation.

LOS CORONADOS ISLANDS

The various authors of this report visited the islands five times between 1935 and 1951. Although the geological study is not yet complete, some of the new information should be presented here. Formation thicknesses are based entirely upon estimates.

Strata of each of the three islands of the Los Coronados group strike almost due north and dip about 25° to the west. Little has been learned of the structural relations of the islands because of the large submerged areas between them; however, faults with throws of several hundred feet occur on Middle and South Islands.

North Island consists of about 500 feet of red to reddish-brown, medium- to coarse-grained sandstone in beds ranging from a few inches to a few feet in thickness. A mineral analysis of three sandstone samples is presented in table 1. Pebbles occur in the lower sandstones but are not common.

Middle Rock is composed of about 125 feet of red to reddish-brown conglomerate, having a sandstone matrix and containing sandstone beds up to 5 feet thick. The sandstone is like that of North Island. Cobbles and boulders of the conglomerate vary between 2 inches and 2 feet in diameter and are chiefly of rounded metavolcanic rocks, schist, and quartzite.

Middle Island is the most complex of the group. The central and eastern part consists of about 50 feet of red conglomerate and sandstone like that of Middle Rock, overlain by about 200 feet of light-green conglomerate and coarse sandstone. The green conglomerate consists of pebbles, cobbles, and boulders of volcanic and metamorphic rocks, including glaucophane schist and others of the Franciscan type. On the northwestern and southwestern sides of the main Middle Island, red conglomerate is in fault contact with the green rocks. A maximum thickness of about 100 feet of the red conglomerate was observed. In each area drag against the fault plane indicates relative upward movement of the red rocks. Thus at least some of the red rocks are older than some of the green ones. An islet east of the main Middle Island, which is separated only at high tide, consists entirely of a third kind of rock, a gray, mildly altered andesite or basalt breccia. The angular clastics range up to 6 feet in diameter. A thickness of about 75 feet of the formation is exposed.

South Island, the largest of the group, consists chiefly of green conglomerate and sandstone like that of Middle Island. A mineral analysis of the sandstone and matrix is presented in table 1. The total thickness appears to be about 700 feet. In the northeastern portion of the island the green conglomerate and sandstone rest upon 120 feet of sandy shales, from which a few Foraminifera were recovered. In fault contact with the sandy shales is a

section of shaly sandstone about 150 feet thick that has moved upward in relative position and may be the lower part of the same formation. This sandstone, in turn, rests upon about 100 feet of green conglomerate and sandstone lithologically like the main formation of the island. The two deep coves on opposite sides of the island are cut in altered volcanic breccia like that of the islet east of Middle Island. In both coves this rock has a thickness of about 125 feet and is overlain by 50–100 feet of sandy shale. The breccia and shale form a fault block that trends across the island and has probably moved upward in the section; thus the breccia is probably the oldest exposed rock of South Island, but its age relation to the red conglomerate and sandstone of Middle and North Islands is unknown.

After a visit to the islands in 1922, Woodford tentatively assigned the green conglomerate of South and Middle Islands to the San Onofre breccia of Miocene age on the basis of its lithology (Woodford, 1925, pp. 182, 183; Beal, 1948, pp. 24, 25). No macrofossils have been discovered in any of the rocks, but we found some Foraminifera in the sandy shale and shaly sandstone in the northeastern part of South Island. They consist entirely of limonitic internal casts, possibly originally of pyrite. Dr. Orville Bandy, of the University of Southern California, identified the following forms: *Bolivina marginata* Cushman, *Buliminella subfusiformis* Cushman, *Nonion costiferus* (Cushman), *Nonionella miocenica* Cushman, (?) *Siphogenerina* sp. (partial internal mold), *Valvulineria* cf. *miocenica* Cushman. These Foraminifera are characteristic of the middle Miocene, supporting the lithologic correlation of the overlying and probably also underlying green conglomerate with the San Onofre breccia.

OFFSHORE

INTRODUCTION

A total of 450 samples containing rocks have been recovered from the offshore area. They can be divided into three classes: bedrock essentially in place, transported rocks, and authigenic rocks. Those essentially in place were distinguished on the basis of angularity, similarity in composition, and fragility (Emery and Shepard, 1945). Hard igneous and metamorphic rocks require much more transportation to become rounded than do friable sandstones, so the former were considered transported, whereas very friable subangular to subround sedimentary rocks were included with the rocks essentially in place. Adjacent samples should show a consistent relationship to each other; a fragment aberrant to the rest of the group is not likely to be in place. Authigenic rocks are confined to phosphorite in its various forms. Detailed descriptions of each sample are not included in this report but are available to the interested reader in a doctoral thesis on the lithology of the area (Butcher, 1951).

ROCKS ESSENTIALLY IN PLACE

General.—Rocks essentially in place occur in three areas of the chart. Two of the areas, near the edge of the shelf, are separated by Coronado Canyon at the 32°30′ parallel of latitude. These are the Coronado Bank area and Los Coronados Islands area. The third area of outcrops extends south from Point Loma for about 5 miles. Table 2 lists the number of samples containing rocks believed to be essentially in place, and figure 4 is a chart showing the distribution of the main types of bedrock.

Coronado Bank.—Proof of the existence of outcrops in the Coronado Bank–

Coronado Escarpment area is obtained from underwater photographs of the bottom. One picture (pl. 1, *A*) reveals the truncated edges of sedimentary beds that form small hogbacks. Dredgings show that the rocks of the bank consist mostly of sandstone and shale (table 2). The sandstones are prevailingly light gray or tan, and in places greenish. Most are fine-grained but some are medium- and coarse-grained; they range from well cemented to poorly cemented and friable. Calcite is generally only a minor constituent, but some beds are cemented with this mineral, and calcareous fossils are not uncommon. Less commonly, limonite serves as the cementing material. The sand grains are mostly angular to subangular and are composed of quartz and feldspar with occasional rock fragments. Biotite is an abundant accessory mineral in some rocks. Many sandstones are silty.

TABLE 2

NUMBER OF SAMPLES CONTAINING ROCKS ESSENTIALLY IN PLACE, CLASSIFIED BY AREA AND TYPE

Rock Type	Coronado Bank	Los Coronados Islands Area	Point Loma Area
Total samples	66	8	44
Sandstone	29	7	15
Shale	40	4	36
Limestone	9	.....	.....
Conglomerate	3	.....	.....

The shales are light gray, light brown, or occasionally greenish gray. Most of them are soft and sandy or silty, but some are black and hard. Calcareous bands and joint fillings occur in some fragments, and fossils are sparse.

White shelly marl, present in small amounts, is rendered friable by abundant silt and sand. Its color ranges from light gray to light brown, depending on the amount of clastic material. The preservation of fossils in the marl is poor; though some Foraminifera, ostracods, echinoid spines and plates, and small corals were found.

A few samples consist of a reddish-brown, poorly sorted conglomerate with round to subround pebbles set in a matrix of angular sand grains. The pebbles are of the metamorphic, intrusive, and volcanic types found in the basement complex to the east. Similar conglomerates probably supplied some of the metamorphic and igneous pebbles that had to be classed as transported because none of the conglomerate matrix remained on them.

Four samples are distinctive enough to warrant separate description. One is a finely laminated, calcareous gray shale, partly weathered to a light brown. The individual laminae are less than 1 mm. thick. Two other samples are of dark-gray shale containing some white, fine-grained sand. The sand is friable and occurs in much distorted beds and even as separate lumps. Possibly it is similar to some of the modern lenses from the deep floor of San Diego Trough described by Ludwick (1950). Lastly, a sample of brown, calcareous, silty sandstone contains two internal casts of a pelecypod identified by Dr. W. H. Popenoe, of the University of California at Los Angeles, as *Thyasira*, which ranges from Oligocene to Recent but is commonest in Pliocene strata.

Foraminifera have been obtained from 13 of the rock samples from the bank. Eleven contain forms which, according to Miss Frances L. Parker, of Scripps Institution, and Dr. Orville Bandy, range from late Pliocene to Recent. These are as follows: *Cassidulina delicata* Cushman,

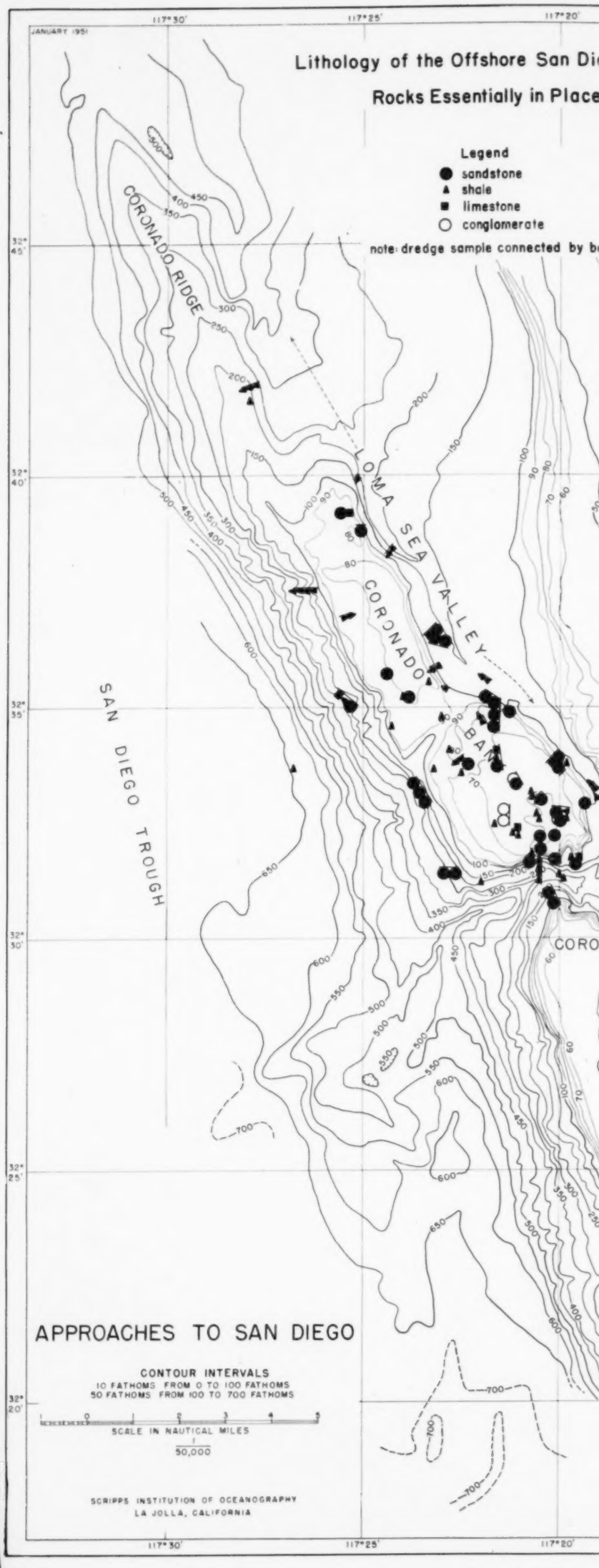
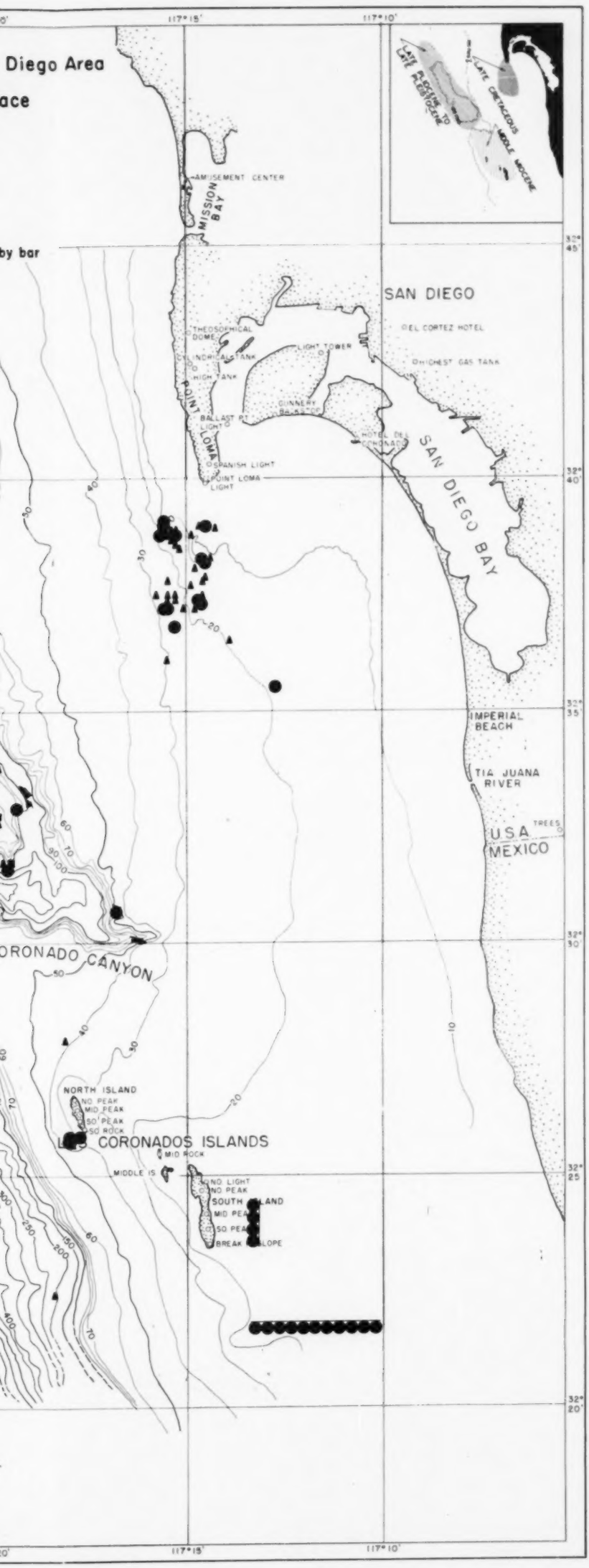


FIG. 4.—Distribution of various



various types of bedrock

C. limbata Cushman and Hughes, *C. tortuosa* Cushman and Hughes, *Bolivina acuminata* Natland, *B. argentea* Cushman, *B. spissa* Cushman, *Uvigerina peregrina* Cushman, *Cibicides fletcheri* Galloway and Wissler, *Rupertia stabilis* Wallich.

If fossiliferous Miocene or Eocene rocks had been eroded to form the bank, reworked Miocene or Eocene Foraminifera might be expected in the unconsolidated sediments. A detailed study of the Foraminifera in 78 sediment samples from the bank showed no forms that do not occur in late Pliocene to Recent formations. Supporting evidence for that age has been obtained from heavy-mineral counts (table 1) on four rock samples. These have appreciable amounts of biotite and actinolite, minerals that have a common association only in the Pliocene to Recent sedimentary rocks of this area (table 1). Glaucophane is present to a limited extent but probably indicates only Miocene or later age.

The minimum age of the rocks would appear to be late Pleistocene, because they are truncated by flat surfaces believed to have been cut by wave action during times of lowered sea-level of the glacial epochs. This physiographic evidence is supplemented by the finding in two sediment samples of shell fragments that have been identified by Dr. Popenoe as not older than Pleistocene. One shell, *Venerupia kennerleyi* (Reeve), is a cold-water form not now known south of latitude 37°. Possibly it lived on the bank during a glacial stage of shallower and colder water.

The San Diego and Sweitzer formations on the adjacent land are also of late Pliocene to Pleistocene age, and, in general, the rocks of the two areas are remarkably similar. The sandstones and conglomerates have the same color in each area, and shell beds are present in both. The formations of the two areas differ in the greater amount of shale and the finer grain size of the sandstones on the bank, but this may have resulted from the greater distance of the present bank area from the shoreline of San Diego–Sweitzer time.

The wide distribution of the samples containing post-late Pliocene fossils and the absence of reworked older fossils indicates that the late Pliocene to late Pleistocene rocks form a continuous blanket of unknown thickness over the bank.

Los Coronados Islands area.—A total of eleven samples, some probably in place, were taken just south of Coronado Canyon and around Los Coronados Islands. Two of these, on the south wall of Coronado Canyon, include large angular pieces of sandstone and shale. The sandstone is yellow to red and coarse-grained. Most of it is lithologically like that on South Island. Two samples from southeast of South Island contain gray, green, and brown fine- to coarse-grained sandstone, which is also similar to rocks of South Island. Two samples from near the south end of North Island are reddish brown and fine-grained, resembling the sandstones of that island. One other, from south of North Island, is light brown, fine-grained, and arkosic, like the sandstone of Middle Island. North of North Island, dredgings produced angular fragments of hard gray shale, grading into a gray, fine-grained silty sandstone. None of these samples is fossiliferous, so their age is not known, but lithologic similarity indicates that they are of the same age as the island rocks.

One sample from the edge of the shelf is a serpentine with slickensides. Although such a rock might be expected along a fault, this sample is probably not in place.

The rocks of the islands have an approximate north strike and a west dip; thus they would lie beneath, and be older than, the rocks of Coronado Bank if there were no intervening faulting or change in strike. No rocks observed on or around the islands resembled those of the bank. Outcrops near the islands should thus be of pre-late Pliocene age. The close resemblance in composition, texture, and color of the Miocene rocks of the islands and those from the surrounding sea floor indicates that the latter are also of Miocene age.

TABLE 3

NUMBER OF SAMPLES CONTAINING TRANSPORTED ROCKS, CLASSIFIED BY AREA AND TYPE

Rock Type	Coronado Bank	Los Coronados Islands Area	Point Loma Area	Tia Juana River Area
Total samples	104	46	76	19
Sandstone	25	14	17	4
Shale	13	1	19	2
Limestone	2			
Conglomerate	1			2
Volcanics	59	30	42	16
Intrusives	11	10	11	5
Metamorphics	22	17	15	8

Point Loma area.—The outcrop area extending southerly from Point Loma consists mostly of shale with a small amount of sandstone, as indicated in table 2. The shale is distinctive: gray to light gray and soft, with a few flakes of biotite. Carbonaceous fragments are quite generally present, and one sample contained a small piece of coal. The shale at this area, unlike the other shales from the sea floor, has a greasy smear when scratched with a knife. The same characteristic is found in some Point Loma shales. It indicates plasticity resulting from a high clay content. Fine-grained, light-gray to grayish-brown, silty, friable sandstones with some biotite are also present in the submarine outcrops. Some of these sandstones are arkosic. Limonitic staining is occasionally present, and greenish staining is less frequent. All these characteristics are similar to those of the Cretaceous Chico formation that crops out on land at the near-by tip of Point Loma.

A heavy-mineral analysis of one of the shales is not very definitive, possibly because of extensive alteration by weathering. Though rare, a few Foraminifera have been recovered from the shale. Dr. O. L. Bandy, of the University of Southern California, identified two species as *Anomolina* sp. and *Cribrostomoides cretacea* aff. Cushman and Goudkoff, which are late Cretaceous in age. The latter is identical with a form found in the Point Loma Cretaceous. Correlation of the submarine outcrops with the Chico of Point Loma is evident.

TRANSPORTED ROCKS

Rock types.—Many of the samples with fragments from outcrops also contained pieces of a transported character. Other samples, however, had transported pieces alone, largely well-rounded metamorphic quartzites and volcanic rocks. Some pebbles of sandstone, shale, limestone, conglomerate, and intrusive rocks were also present. Table 3 lists the transported samples according to their gross lithology. Figure 5 shows that most of the transported rocks occur in the same areas that have outcrops: the Coronado Bank, Los Coronados Islands, and the Point Loma areas. There is an additional large area of transported samples near the coast off the Tia Juana River mouth, a smaller area farther north along Coronado Strand, and a still smaller one in the entrance to San Diego Bay. Some

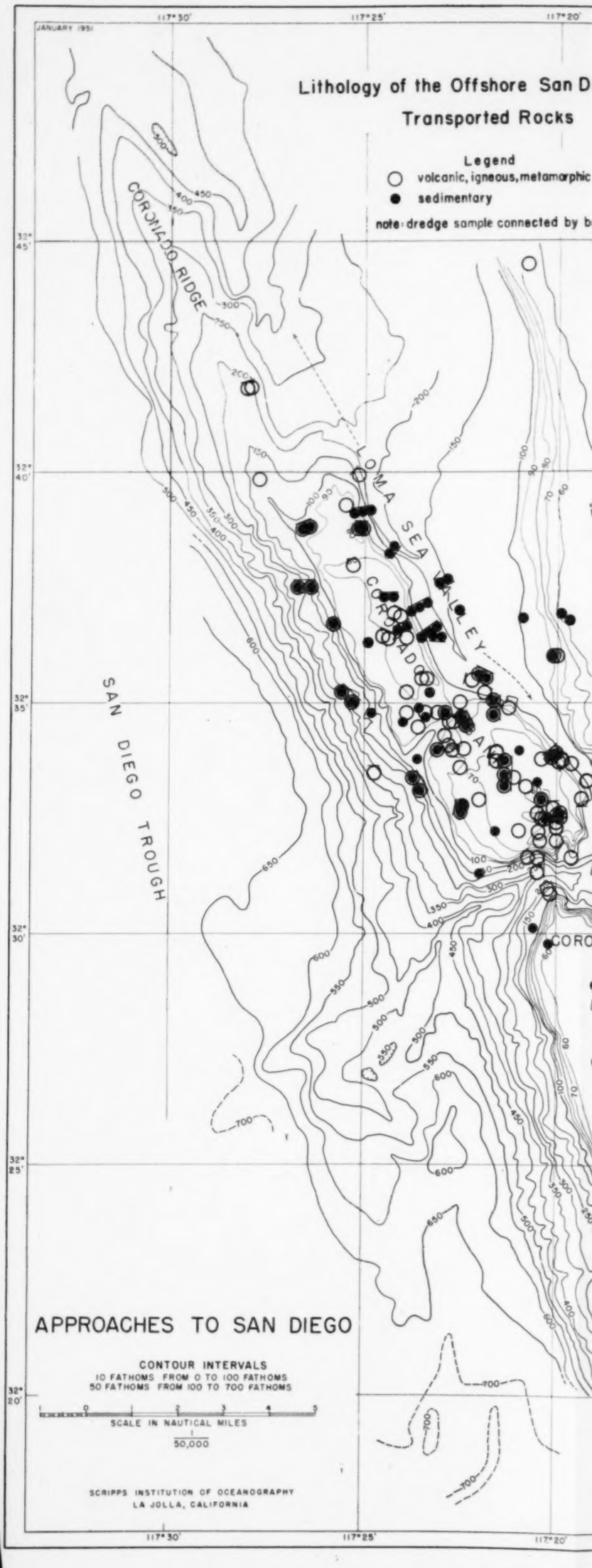
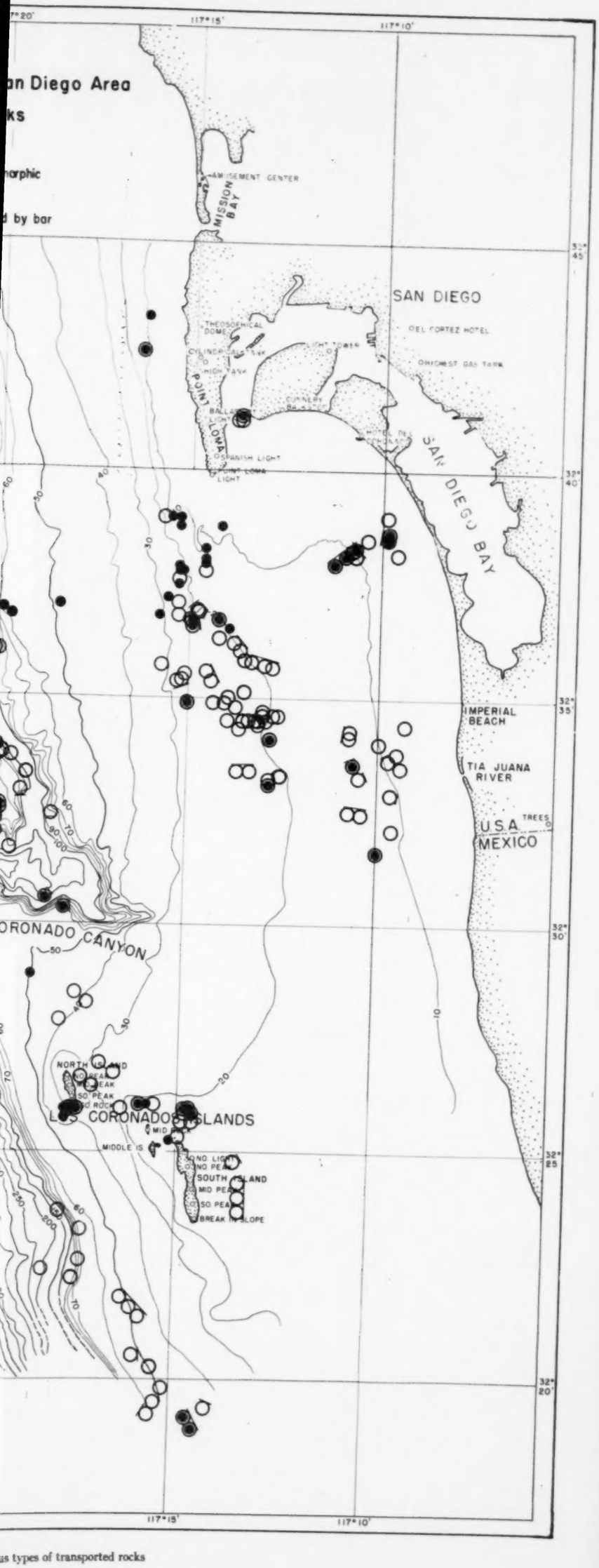


FIG. 5.—Distribution of various t



as types of transported rocks

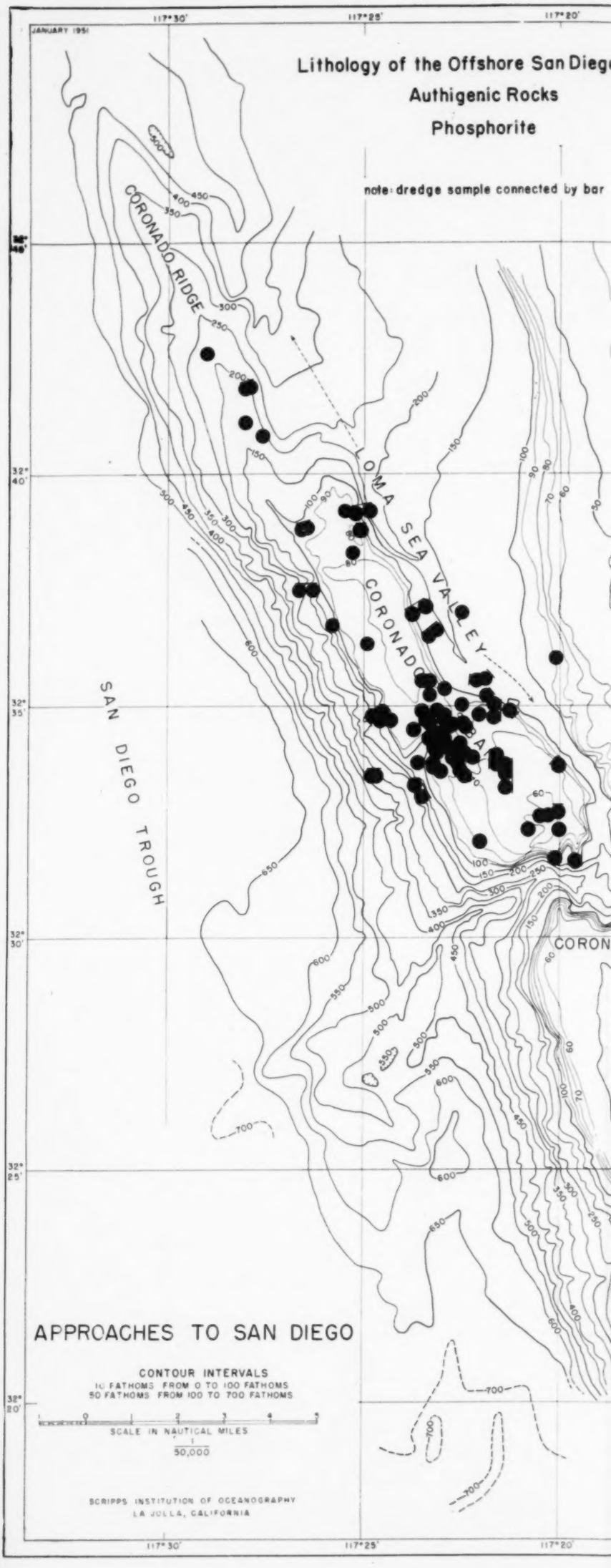
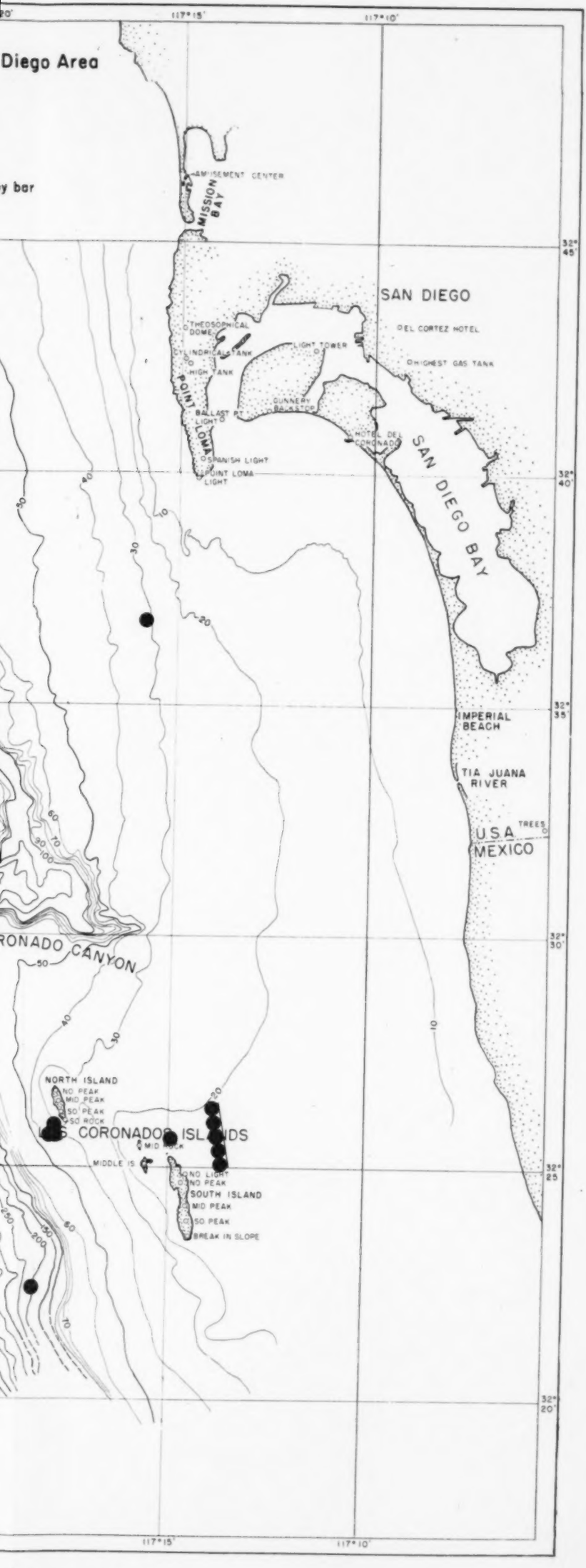


FIG. 6.—Distribution of phosphorite,



ite, an authigenic rock

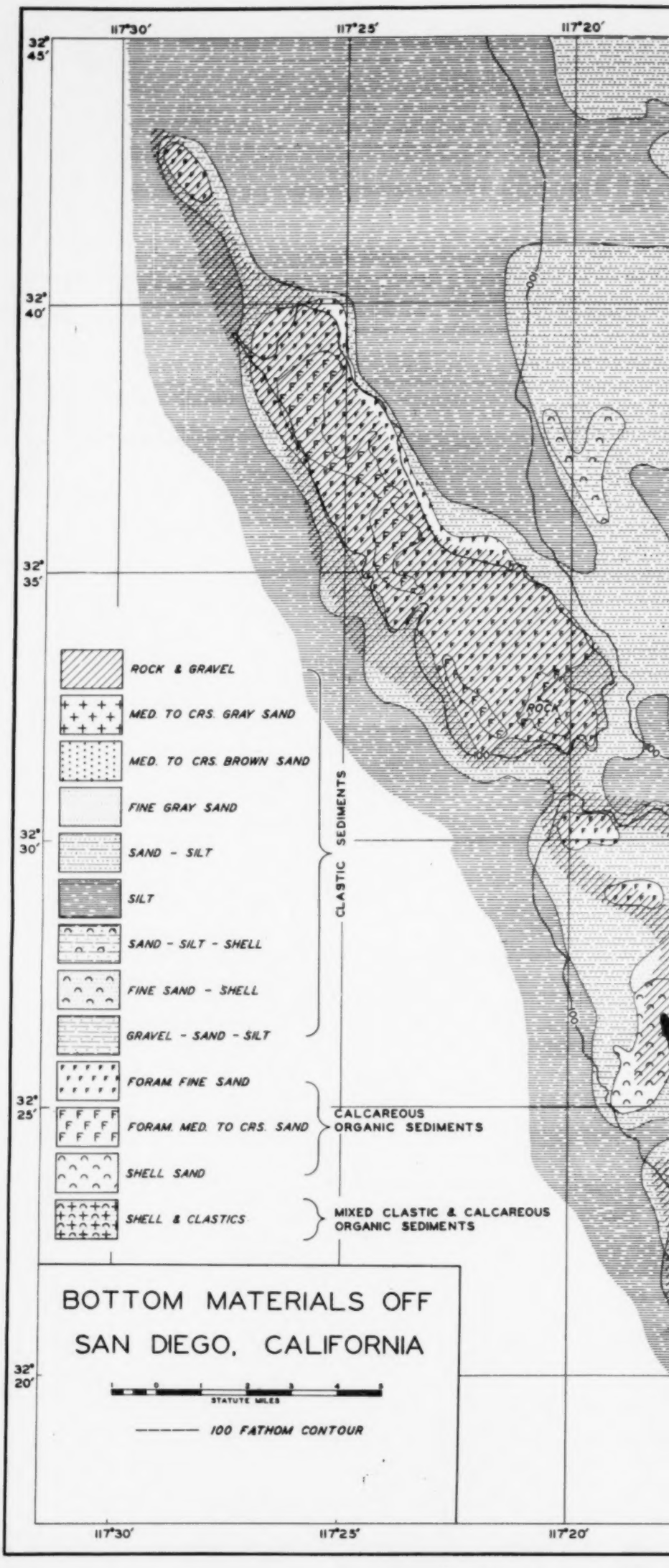
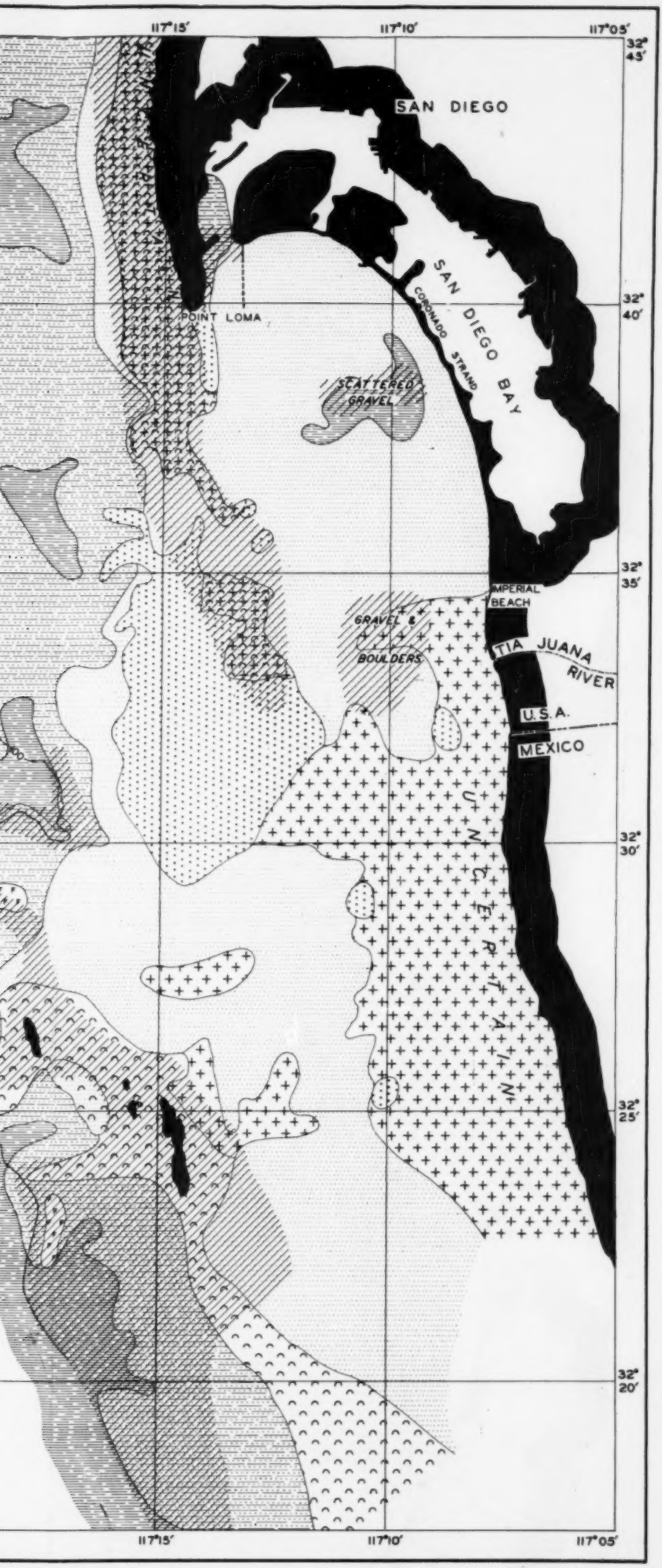


FIG. 7.—Classification and distributio



ibution of unconsolidated sediments

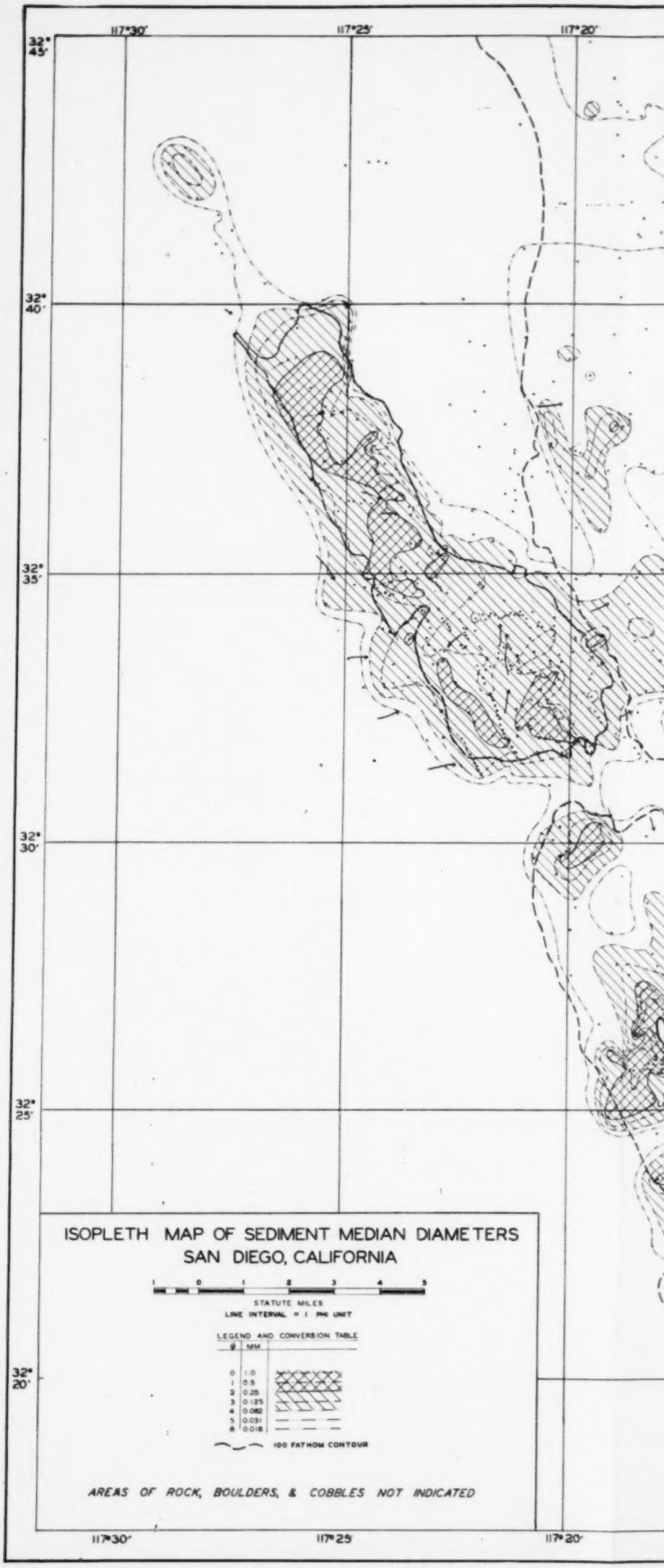
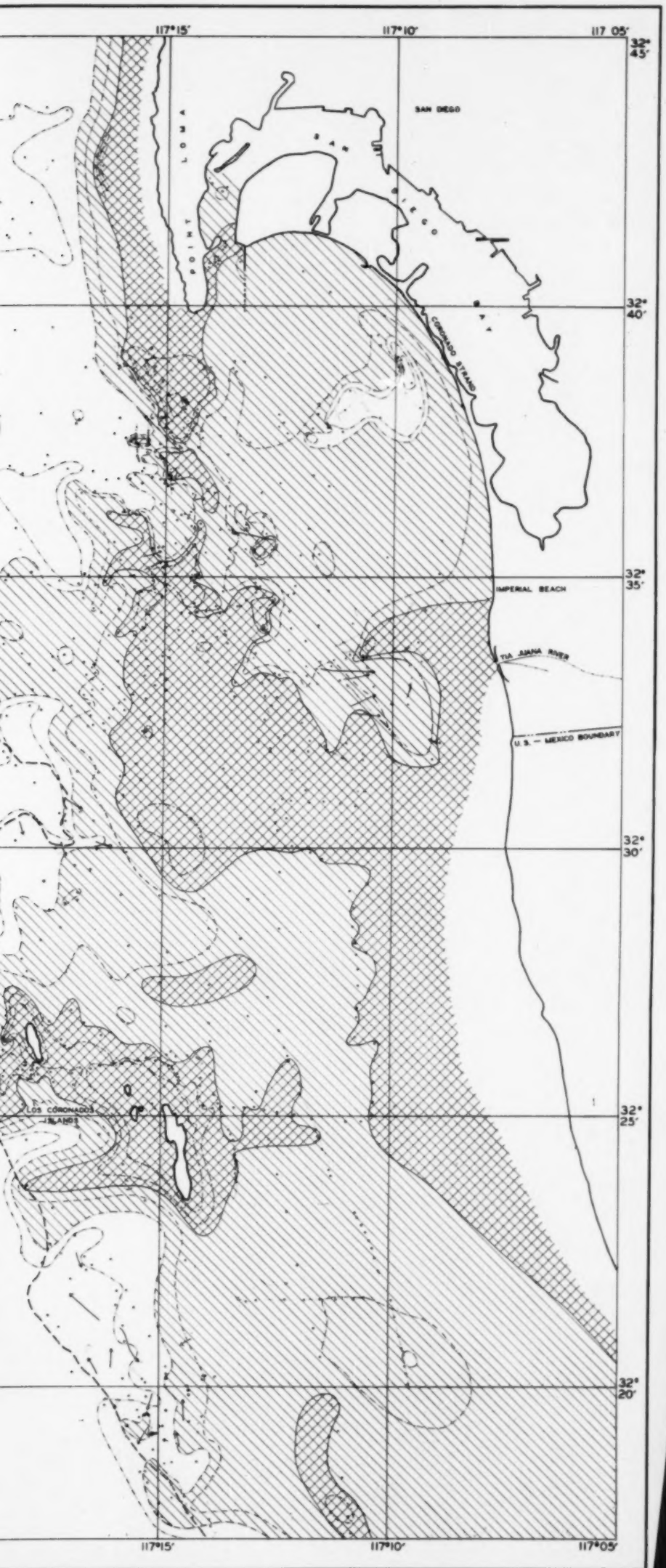


FIG. 8.—Isopleth chart of median diameters of unconsolidated sediments off San Diego. In some



some areas which are dominantly rocks or boulders, the sand represented here is only interstitial

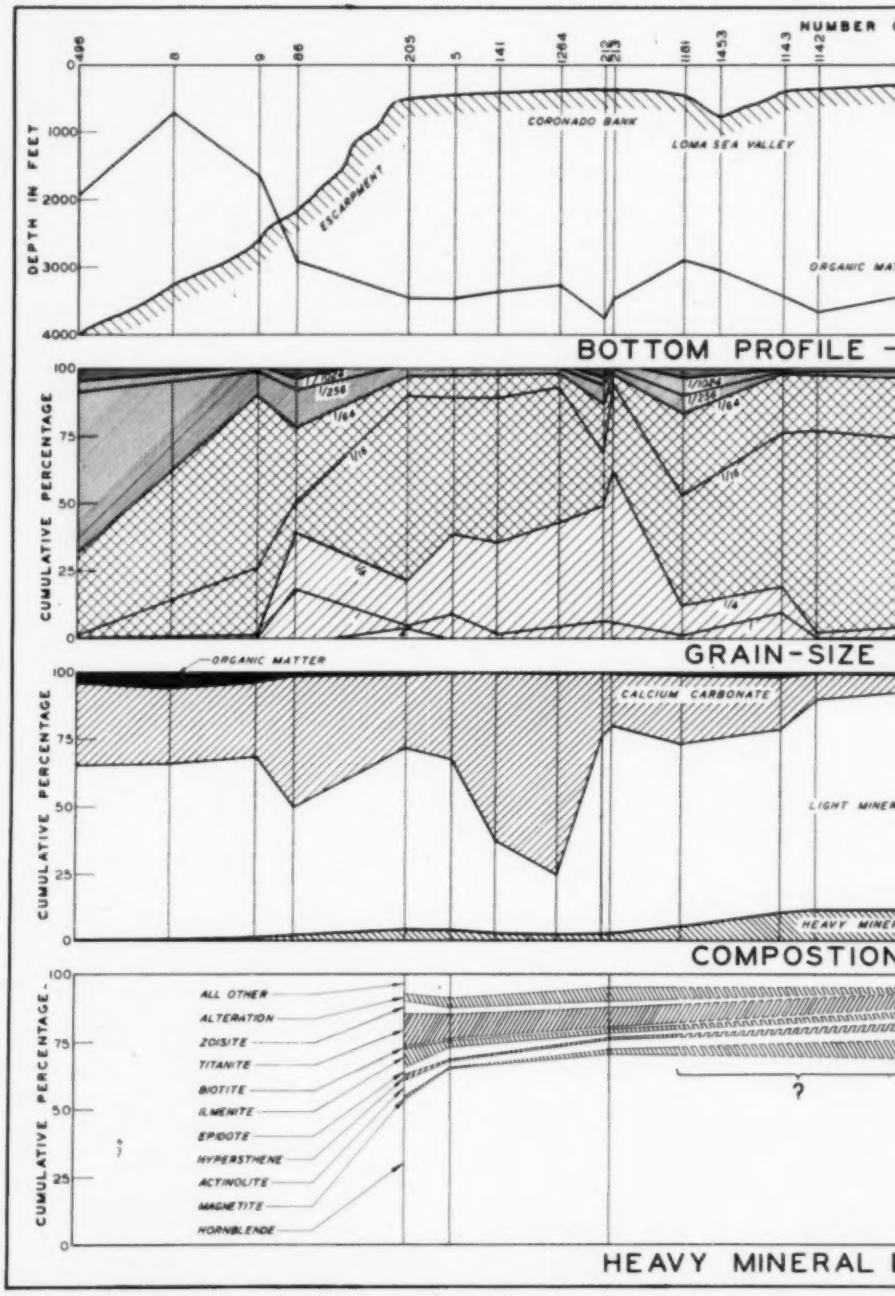
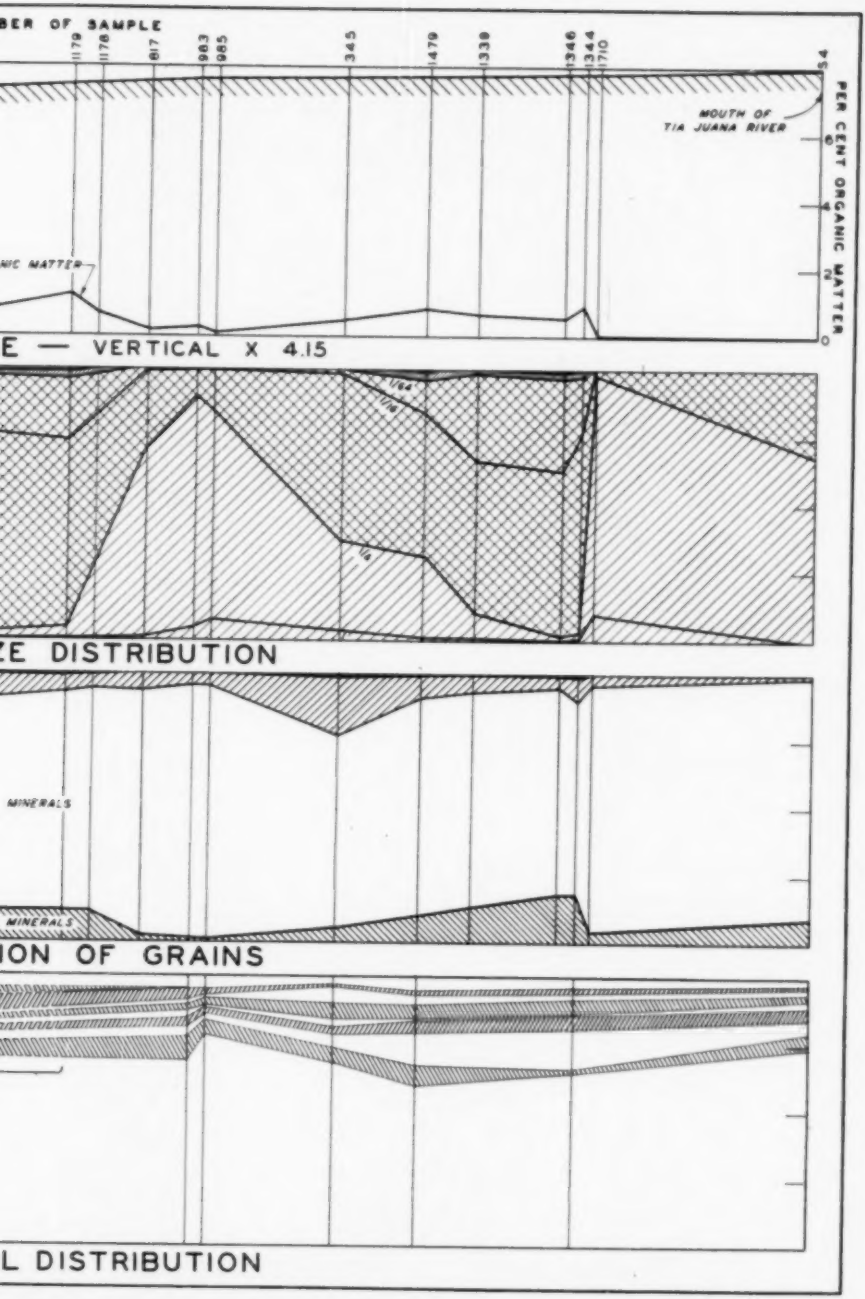


FIG. 10.—Cross section from mouth of Tia Juana River westward across shelf, Loma Sea Valley, Coronado B
in samples along the section are shown.



ado Bank, and top of Coronado Escarpment. The grain size, general composition, and chief heavy minerals

transported rocks also occur along the axis of Loma Sea Valley.

Sources.—The transported rocks of the submarine section must have been derived from outcrops in the area, from present or past land areas, or from some combination of both. Metamorphic, intrusive, or volcanic rocks occur in 65–80 per cent of the samples in each area. The remainder of the samples consists only of sedimentary rocks. If the transported rocks were derived entirely from near-by outcrops, the lithologic types should be similar to those of the bedrock, but not exactly in the same ratios because of differences in hardness, the more easily abraded rocks undergoing the greater loss during transportation. For example, sandstone and shale occur in a ratio of 3 to 4 in the bedrock of the Coronado Bank area, and in the transported rocks of the same area they occur in a ratio of 2 to 1. In the bedrock of the Point Loma area, shale is more than twice as abundant as sandstone, and in the transported rocks of the same area shale and sandstone are about equal.

Had the transported materials been derived from a distant source and the distributing agencies acted over the whole area, the same ratios of lithologic types should exist over the whole area. Thus derivation from the east and redistribution by waves or streams should produce a uniformity of composition in the transported materials. Table 3 shows that the volcanic, intrusive, and metamorphic rocks do have a nearly constant ratio to one another. If these rocks were derived only from local sources, none should occur in the Point Loma area, where there are no conglomerates in the submarine section to provide the crystalline rocks. It seems probable that most of the volcanic, metamorphic, and intrusive rocks have been derived from distant sources and spread over the entire area by stream or marine agencies.

The large area of rounded cobbles and boulders off the Tia Juana River is noteworthy for an almost complete lack of fine sediment cover and for the absence of bedrock. The rounded rocks from that area are almost exclusively of intrusive, metamorphic, and volcanic types. Encrustations on the rocks show that they are not now being moved about. Probably they were deposited by the river sometime in the past. The rock types in this area may well be indicative of the kinds that were transported from land and deposited with locally derived rocks in the other areas of the sea floor. The boulder area off Coronado Strand farther north may be of a different origin because of the predominance in it of sedimentary rocks, though possibly the latter were added by kelp-rafting of rocks from the up-current Point Loma area, where kelp is abundant. Gravel in the entrance of San Diego Bay may be ballast that was dumped by sailing ships during the last century. This is said to be the origin of the near-by Ballast Point.

If gravels were deposited atop bedrock over the whole shelf and bank, subsequent deposition of sands and finer sediments has buried them except where currents are strong enough to by-pass the fine-grained sediments, leaving the bedrock and scattered gravels uncovered. The Coronado Bank and Los Coronados Islands areas lie at the edge of the shelf, where tidal and other currents are accentuated. The area of rock and cobbles extending southerly from Point Loma probably results from strong coastal currents, driven by northwesterly winds and leaving the coast at the tip of Point Loma. In addition, cobbles unrelated to bedrock occur in the entrance to San

Diego Bay, where tidal currents keep them exposed, and in two other areas near the coast and farther south. No explanation for lack of burial of the latter areas can be given.

Date of deposition.—Several lines of evidence suggest that the gravels were deposited during one of the glacial times of lowered sea-level. The bedrock on which they rest contains late Pliocene to Recent fossils, and the gravels are, of course, younger than the rock. Because the boulders rest on the flattened 55-fathom surface of Coronado Bank, they must have been deposited after or during the formation of that surface. Flint (1949, p. 435) speculated that benching at such a depth may have occurred during the last glacial lowering of sea-level. One sample contains a 56-cm. limestone boulder, and another a 36-cm. andesite boulder. These sizes are probably too large for kelp-rafting, and, so far as can be determined, bottom currents are not strong enough at the present time either to transport the cobbles and boulders or to erode the bedrock to provide materials of local derivation. Abundant encrusting organisms on the exposed surfaces show not only that the gravels are not being moved now but also that effective scour by interstitial sand does not occur. Currents sufficiently strong for such transport or erosion probably have not occurred since the period of relatively lowered sea-level during the last glacial stage. Dietz and Menard (1951) have pointed out that active erosion of the sea bottom by waves is restricted to depths of less than about 5 fathoms and that most of the area of continental shelves is now a deeply submerged remnant of marine erosion during times of glacially lowered sea-level. This conclusion appears to be supported by the great abundance of empty pholad borings in the rocks. Ricketts and Calvin (1948, p. 178) state that, although the ecology of pholads is imperfectly known, they are apparently limited in their greatest abundance to the intertidal zone, though occasionally occurring in dredged rocks as deep as 30 fathoms. The abundance of the borings suggests that they were made at a time of lowered sea-level and that the rocks have subsequently remained undisturbed. Coarse terrace deposits also characterize the Pleistocene formations on land; hence transported materials were evidently spread over the whole area during the Pleistocene epoch.

AUTHIGENIC ROCKS, PHOSPHORITE

Characteristics.—The only authigenic rock is phosphorite. This was found in 83 samples from Coronado Bank and in only 8 samples from other areas. Figure 6 shows the distribution of these samples.

The phosphorite ranges in size from a few millimeters to 43 cm. Much of it is nodular and polished as though in the process of formation. Most of the phosphorite contains the usual irregular layers that indicate discontinuous growth, but some is aphanitic and structureless. Broken and re-cemented pieces occur occasionally, and several nodular specimens include pebbles of clastic material. Some phosphatic particles contain so many silt and sand grains that they are difficult to distinguish from non-phosphatic rocks. Films of phosphorite also coat many of the other rocks of the bank, so that rocks must be broken open to determine their composition.

Fossils are not rare but are difficult to separate for identification. Borings in the phosphorite are filled with modern sediments and Foraminifera. Light-brown material is present on the outer parts of the nodules, as mentioned by Dietz, Emery, and Shepard (1942, p. 827), and

dark-brown to black phosphorite generally occurs in the center of the nodules or as fragments cemented together by the lighter-brown phosphorite. The Foraminifera from these two types of phosphorite seem to show no difference in age, unlike the phosphorite dredged elsewhere off southern California. Foraminifera have been obtained from six samples. Dr. Orville Bandy identified the following forms: *Cassidulina limbata* Cushman and Hughes, *C. tortuosa* Cushman and Hughes, *C. cf. californica* Cushman and Hughes, *Cibicides fletcheri* Galloway and Wissler, *Rupertia stabilis* Wallich. This fauna is similar to that obtained from the underlying rocks, except for the absence of *Bolivina* and *Uvigerina* in the phosphorite. Reference to Natland's (1933) chart of the temperature and depth distribution of Foraminifera off California suggests that the Foraminifera of the phosphorite are characteristic of slightly shallower depth (his zone III, 125–900 feet) or slightly younger age than those of the underlying rock. In addition, two of the phosphorite nodules include a poorly preserved foraminiferal fauna possibly of Miocene age, including *Cassidulina crassa* d'Orbigny, mixed with late Pliocene to Recent species. They may well have been redeposited from Miocene rocks of Los Coronados Islands.

In situ origin.—The deposition *in situ* suggested by Dietz, Emery, and Shepard (1942) is borne out in the examination of the samples of this area by (1) the lack of abrasional rounding, (2) the fresh nodular appearance of most of the surfaces, (3) the phosphatic coatings on many boulders, and (4) the included sediment, which is of the same type as that which surrounds the nodules. Some nodules show an irregular concentric layering of phosphate and thin manganese oxide. In these nodules the direction of major growth is toward the side covered with living organisms, indicating growth in the present position. Deposition *in situ* is also suggested by the localization of the deposit; transportation with other rocks should have left deposits in other parts of the sea floor. Furthermore, no phosphorite has been reported in the rocks exposed on the near-by land. The mode of formation is apparently by chemical precipitation, although proof of this rests on further investigation of the physical-chemical properties of sea water with respect to phosphate deposition.

Date of deposition.—Dietz, Emery, and Shepard (1942) suggested that the phosphorite off southern California is largely Pliocene or Quaternary in age, though Miocene Foraminifera were most commonly included. Later, Emery and Dietz (1950) gave evidence for a Miocene age for the bulk of the phosphorite. A Quaternary age is indicated for the phosphorite off San Diego because of the enclosed Foraminifera and the stratigraphic relations. The phosphorite rests in part upon a surface of the bank that ranges from 75 to 90 fathoms in depth and has a slope to the north and west from the shallow southern end of the bank. If the phosphorite were formed *in situ*, it must be younger than this surface. Because the rocks forming the surface are late Pliocene at the oldest, the phosphorite must be younger than late Pliocene. The surface at about 55 fathoms of presumably late glacial age has practically no phosphorite nodules. Only one nodule was found within the 60-fathom contour of figure 1. Therefore, most of the phosphorite appears to be older than the last glacial stage and, thus, is pre-late Pleistocene. A similar relationship exists at Fortymile Bank farther offshore (Emery and Dietz, 1950).

Eight samples containing phosphorite

were obtained outside the bank area. One was found off Point Loma, where the water is shallower and the sediment less fossiliferous than on the bank. One came from the east side of Loma Sea Valley near Coronado Bank. The remaining six samples are from the area surrounding Los Coronados Islands. The environment of these latter samples is similar to that of the bank, for both the depths and the sediments are comparable. The phosphorite of three of these samples replaces shells, none of which are larger than 5 cm. Possibly they represent a later deposition of phosphorite than the large nodules of Coronado Bank and can be correlated with the coatings found on other kinds of rock.

Unconsolidated Sediments

General

Knowledge of the unconsolidated sediments is based on a study of more than thirteen hundred samples from all parts of the submarine area. The coarse unconsolidated materials, such as granules, pebbles, cobbles, and boulders, have been discussed in the section on lithology, because of their close association with the underlying bedrock. There is reason to believe that some of the sediments constituting the surface layer were laid down when the sea-level was relatively lower than now and are consequently older than the materials now being deposited. Since definite age determinations are lacking, however, we have included all the finer surficial materials in the discussion of unconsolidated sediments.

Classified according to composition, the modern sediments fall into three broad groups: (1) clastic sediments, (2) calcareous organic sediments, and (3) mixed clastic and calcareous organic sediments. Strictly speaking, neither the clastic nor the organic sediments are wholly of clastic or organic origin. However, the clastic sediments consist predominantly of detrital materials and the organic sediments predominantly of calcareous organisms. The third group, mixed clastic and calcareous organic sediments, contains approximately equal amounts of the two main types. Of these three broad groups, the sediments of clastic origin are by far the most widespread, covering 685 square statute miles, approximately 90 per cent of the whole submarine area. In contrast, the calcareous organic sediments make up only 8 per cent, or 68 square miles. Least abundant are the mixed sediments, which cover only 2 per cent, or 15 square miles.

These three major sediment groups have been further subdivided on the basis of texture and color. Altogether, 13 sediment types have been distinguished. Of this number, 9 are clastic sediments, 3 are calcareous organic sediments, and 1 is of mixed origin. The areas covered by each type of sediment are shown in chart form in figure 7. In the areas of bedrock and gravels, indicated by shading, sediment forms patches on the rock and lies between the boulders and cobbles.

Areal Distribution and Description of Sediment Types

Clastic Sediments

Clastic sediments cover most of the shelf and all the area of the submarine slopes and valley floors. These sediments can be subdivided into five major types and four minor ones (fig. 7).

Medium to coarse gray sand.—The sediment consists of medium to coarse, light to medium olive-gray sand at Imperial Beach, in two isolated patches east and north of South Coronado Island, and in a

large area off the mouth of the Tia Juana River and the Mexican coast. The area of sand off the Mexican coast may be less extensive than is shown in figure 7, because few samples are available there. The sand of Imperial Beach and the sand of the two isolated patches have a salt-and-pepper appearance and are slightly lighter in color than that off the river and the Mexican coast. This difference results from a gray coating on the sand grains in the latter area. The grains in all these areas are predominantly angular to subangular.

Medium to coarse brown sand.—A large area of medium to coarse light-reddish-brown sand occurs west of the medium to coarse gray sand and approximately in line with the Tia Juana River. The sand in this zone is notably coarser along its western boundary than elsewhere. Brown sand also was found in two small isolated patches at the outer edge of the large area of medium to coarse gray sand and in an elongate patch just south of Point Loma. Reddish sand, possibly the same as the submarine brown sand, occurs in some well borings near the mouth of the Tia Juana River. The brown sand may once have formed a practically continuous layer from Point Loma on the north to its present southern boundary at latitude 32°25′, the present discontinuity resulting from partial burial beneath medium to coarse gray sand and fine gray sand.

The brown and gray sands are remarkably similar in grain size, shape, and mineralogy. Grains of the brown sand are stained with yellow and red iron oxides. Ferric iron in the stain, determined for 3 samples of coarse brown sand, averages 0.41 per cent, and for 3 samples of coarse gray sand, only 0.25 per cent. Qualitative tests for ferrous iron were negative or weak.

Fine gray sand.—The most widespread of the clastic sediments on the shelf is a fine, light-olive-gray sand (fig. 7), which generally does not extend seaward of the 40-fathom contour. This sand is separated into a northern and a southern zone by a discontinuous arc of other bottom materials between the Tia Juana River and Point Loma. A narrow fringe of the fine gray sand also occurs along the western boundaries of the main brown-sand area and of the shell and gravel area west of Point Loma.

The northern fine-gray-sand zone consists chiefly of fine and very fine sand. The beach presents a gradation from medium and coarse sand at Imperial Beach to predominantly fine sand along Coronado Strand. This beach sand has a salt-and-pepper appearance and is slightly lighter in color than the submarine sand. Near shore the principal constituent is fine sand with minor amounts of medium sand. To the south and west there is a gradation into very fine sand with an appreciable quantity of silt. Samples along the southern and western margins of the zone show some mixture of the fine gray sand with the bordering coarse brown and gray sands.

The sediment of the southern fine-gray-sand zone consists predominantly of fine sand, with a considerably larger proportion of medium sand than is present in the northern zone. It is also noticeably lighter in color than the sand of the northern zone, and it has a salt-and-pepper appearance like the beach sand from Coronado Strand and the medium and coarse gray sand on the shelf east of Los Coronados Islands.

The fringe of fine gray sand along the western border of the main area of coarse brown sand and the area of shell and gravel farther north consists primarily of very fine sand with appreciable amounts

of silt. Westward it grades within a short distance into the sand-silt area. The grains of the fine gray sand are predominantly angular to subangular.

Gray sand-silt.—The sediments of the outer edge of the shelf and the upper portions of the submarine slopes bounding San Diego Trough, Loma Sea Valley, and Coronado Canyon contain approximately equal proportions of very fine light-olive-gray sand and silt. Only a minor amount of clay is present. This sand-silt zone extends shoreward of the 30-fathom contour only south of Los Coronados Islands, but it reaches 250 fathoms in Coronado Canyon. In the Coronado Bank area, the sand-silt forms a narrow fringe surrounding the shallower calcareous organic sediments, and in turn the sand-silt grades into the silt of deeper water.

Coprolites are common constituents of the sand-silt, and the tests of Foraminifera and fragments of mollusks and other organisms are fairly abundant. Most of the detrital grains are angular, and many are coated with a light-green to grayish-brown stain.

Gray silt.—In the deeper parts of the offshore area (Loma Sea Valley and San Diego Trough) and beyond the sand-silt area, the sediment is a pale olive-gray silt with minor amounts (less than 15 per cent) of very fine sand and clay. Isolated patches of this material also occur on the floor of Coronado Canyon near its head and on the outer edge of the shelf in the vicinity of latitude 32°36′. The gray silt is the most extensive of all the unconsolidated sediments, covering about 350 square statute miles (46 per cent) of the total submarine area. Most of the coarser constituents are angular and slightly coated with a green to gray-brown stain.

Minor sediment types.—Among the clastic sediments are several types, of small areal extent, that do not belong to the sediments already discussed. The first of these is a light-olive-gray sand–silt–shell sediment which occurs on the eastern slope of Loma Sea Valley in the vicinity of latitude 32°37′. It differs from the surrounding sand-silt chiefly in its higher percentage of coarse shell fragments.

Just west of Middle Coronado Island is another sediment type similar to the fine gray sand to the north but containing a fairly high percentage of shell fragments.

A third minor type of clastic sediment is a very light-gray mud, composed predominantly of silt and clay (shown as silt in fig. 7). This was found during World War II in an isolated patch associated with boulders off Coronado Strand at latitude 32°38′. It resulted from the pumping of fine muds from the floor of San Diego Bay over Coronado Strand to the open sea in 1942, just prior to the collection of samples. By 1950, wave and current action had removed these fine sediments.

Another local type occurs along the east side of Point Loma and in the entrance of San Diego Bay. This consists of poorly sorted gravel-sand-silt.

CALCAREOUS ORGANIC SEDIMENTS

General.—Although the calcareous organic sediments cover only 8 per cent of the offshore area, they are in several respects the most interesting of all the modern sediments. Perhaps the most striking feature is the general correspondence of their areal distribution to outcrops of bedrock and deposits of boulders and cobbles (figs. 4 and 7). Thus they occur only in areas where clastic sediments are not accumulating now or where their rate of accumulation is exceedingly slow.

The organic sediments are limited to Coronado Bank and to the shelf west and south of Los Coronados Islands. They generally contain appreciable quantities of nonorganic material, which in some places is evidently detrital but which elsewhere appears to be residual from the underlying bedrock upon which the organic sediments rest. Classified on the basis of composition and texture, these sediments are divided into three types.

Foraminiferal fine sand.—The organic sediment of greatest areal extent is a fine pale-olive foraminiferal sand. This material covers most of the top of Coronado Bank and is present in one isolated patch on the ridge north of Coronado Bank and in three small areas near the edge of the shelf south of Coronado Canyon. Samples consist principally of tests of planktonic and benthonic Foraminifera and contain a smaller but appreciable quantity of shell fragments, glauconite, and coprolites. Clay casts in many of the foraminiferal tests are also abundant in the bank sediments. They will be discussed in more detail in a later section of this report.

Detrital grains range from very little to more than half of the material. They vary from coarse sand to silt, grading into the finer sizes near the boundary between the foraminiferal sand and the sand-silt areas. Most of the detrital grains are subangular to subround and are coated with a light greenish-gray stain.

Foraminiferal medium to coarse sand.—A pale-olive foraminiferal medium to coarse sand occurs on the shoaler portions of Coronado Bank. This is similar to the foraminiferal fine sand except for the coarser texture of the detrital grains. In many samples detrital grains of coarse and medium sand make up approximately 50 per cent of the material. Most of them are subround, but some are subangular to angular. Many are coated with a light-gray stain.

Shell sand.—The third type of calcareous organic sediment is a very pale-brown to light-grayish-brown coarse shell sand. This is confined principally to the area surrounding Los Coronados Islands and to a narrow strip of the shelf between the 20- and 30-fathom contours and trending southeasterly from the islands. Throughout most of this area the sediment consists almost entirely of coarse shell fragments, but near the islands the shell sand is mixed with gravel, sand, and silt in varying proportions. These detrital grains are mostly subangular to subround and are coated with a yellowish-brown to gray stain. In contrast, the detrital grains near the boundary of the shell-sand area are similar to those of the adjacent clastic sediments, being mostly angular to subangular and only slightly stained.

MIXED CLASTIC AND CALCAREOUS ORGANIC SEDIMENTS

The sediments that consist of approximately equal proportions of calcareous organic and clastic constituents have the smallest areal extent of the three groups of unconsolidated materials. They are confined mainly to a narrow strip bordering the west side of Point Loma and extending for approximately 4 miles south of the tip of Point Loma (fig. 7). About a mile farther south is a small, isolated area of similar material. Possibly these areas were formerly continuous, and their juncture has been obscured by a cover of fine gray sand. The continuity of the mixed sediments near the west side of Point Loma may not be correct, as shown in figure 7, because of the scarcity of near-shore samples in that area.

The mixed sediments range from very

light gray to medium olive-gray and consist of approximately equal amounts of coarse broken shells and of granules and pebbles. Near its boundary with the area of fine gray sand the mixed sediment contains some fine sand and silt.

TEXTURE

GRAIN SIZE

Enough sediment for satisfactory analysis was present in 1,225 of the 1,300 samples available. So much time would have been required to make conventional mechanical analyses of so many samples that a rapid visual comparison method was developed (Emery and Gould, 1948). To serve as standards, 166 of the samples were analyzed by the usual screening and pipetting method.

The median diameters of all 1,225 samples were found to range from 0.011 to 2.0 mm. All analyses included both clastic and calcareous grains but excluded gravel. If gravel had been included in the analyses, larger medians would have resulted for some samples. Nearly 50 per cent of the samples have medians within one ϕ unit (one Wentworth grade size—see fig. 8) of 0.125 mm., thus being classified as fine sand or very fine sand. An approximation of the over-all size distribution of the sediment was obtained from the weight percentage in each size grade of the 141 conventionally analyzed samples from the shelf shallower than 100 fathoms and of the 25 samples from deeper water. These were totaled and plotted as though for two composite samples. The resulting plot (fig. 9) shows that the median diameter of the composite sediment from the shelf and bank is 0.137 mm., whereas that of the deeper sediment is only 0.038 mm.

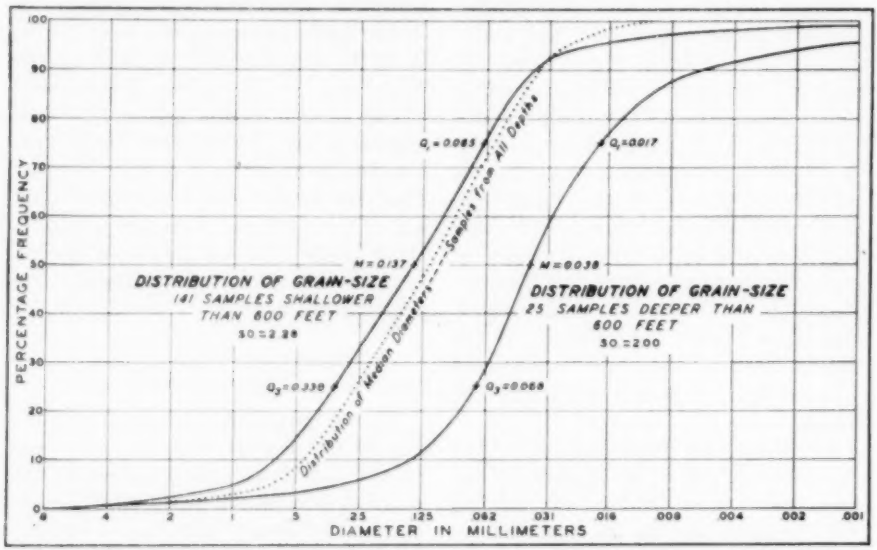


FIG. 9.—Distribution of grain sizes in an average sample from the shelf and bank top (less than 100 fathoms) and from deeper water (more than 100 fathoms). Both were compiled by averaging the weight percentage of each grade size in all samples. For comparison, the distribution of median diameters of samples from all depths is shown by a dotted line.

The medians of all samples were plotted on a map at the positions of the sampling sites. Isopleth lines at one ϕ unit intervals were drawn and shaded by cross-hatching. The resulting figure 8 shows that the coarsest sand (-2–0ϕ, or 4–1 mm.) is close to the cliffed shores of Los Coronados Islands. Intermediate sand (0–2ϕ, or 1–$\frac{1}{4}$ mm.) forms a broad zone along the coast, except off Coronado Strand, where it is separated from the shore by a large zone of finer sand. It also borders the coarse sand of Los Coronados Islands and occurs atop Coronado Bank. Finer sand (2–4ϕ, or $\frac{1}{4}$–$\frac{1}{16}$ mm.) mainly occupies a wide area off Coronado Strand and occurs also in strips fringing the intermediate sand. The fourth main category of sediment, silt (4–6ϕ, or $\frac{1}{16}$–$\frac{1}{64}$ mm.) occurs chiefly in the deeper water of the bottom and side of San Diego Trough, in Loma Sea Valley, and at the head of Coronado Canyon. The finest sediment of the entire region was discovered outside Coronado Strand in depths of only 6–9 fathoms and fully exposed to the waves. As previously stated, this anomalous occurrence resulted from the pumping of fine muds from the floor of San Diego Bay over the strand to the open sea.

The distribution of grain sizes from the shoreline to the Coronado escarpment is illustrated by a cross section in figure 10. This clearly shows the patchy areas of coarse and fine sand on the shelf, the fine sediment in Loma Sea Valley, the coarsening on Coronado Bank, and the marked decrease in grain size on the escarpment leading down to the San Diego Trough. Exposed bedrock and gravel were omitted, to simplify the drawing.

SORTING

Sorting coefficients for conventionally analyzed samples range from 1.18 to 5.56. As shown by figure 11, this distribution is such that 95 per cent of the samples have sorting coefficients smaller than 2.5 and, according to Trask's limits, would therefore be considered well sorted. Sorting coefficients of the individual samples are widely scattered with respect to grain size, but reference to figure 9 shows that the theoretical composite sample from depths of less than 100 fathoms and the one from deeper water have coefficients of 2.26 and 2.00, respectively. These values are in the reverse order of what might be expected from Inman's (1949) analysis of sorting coefficients in the light of fluid mechanics, wherein he found the best sorting for sediment of 0.18 mm. median diameter and progressively poorer sorting for coarser and finer sediments. However, his study was necessarily based on detrital grains that are in equilibrium with the currents, in contrast to the shelf sediments that are a composite of detrital, residual, organic, and authigenic grains, mostly too coarse to be affected by present-day currents. Had mechanical analyses been made for the insoluble residues, the shelf and bank sediment would have been coarser and better sorted than the sediment from deeper than 100 fathoms. Comparison of the actual relationship of sorting and grain size with the equilibrium relation-

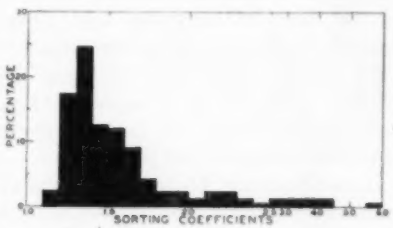


FIG. 11.—Histogram showing variation of sorting coefficients of 158 unconsolidated sediment samples in all depths.

ship thus offers a means of judging whether the detrital fractions were deposited under present or past conditions.

ORGANIC CONSTITUENTS

CALCIUM CARBONATE

The percentage by dry weight of material soluble in dilute hydrochloric acid was determined for 108 samples. The soluble fraction is termed "calcium carbonate," though it contains, in addition, small quantities of magnesium carbonate, collophane, and other compounds. The bulk of this "calcium carbonate" is in the form of benthonic and pelagic foraminiferal tests and broken fragments of pelecypod shells and bryozoans. The median value for all samples is 11 per cent, whereas that for samples deeper than 100 fathoms is 21 per cent.

Isosol lines (fig. 12) were drawn for the area on the basis of 108 analyses aided by visual comparison of numerous other samples with the analyzed set. The isosol lines were chosen on a logarithmic progression (5, 10, 20, 40, and 80 per cent) because of the wide range of values (0.8–86.7 per cent). Owing to the small number of analyses, the lines are broadly curving. A denser control would have revealed numerous irregularities and obscured the simple general picture. The chart (fig. 12) shows that less than 1 per cent soluble material occurs in the beach sands of Coronado Strand and less than 5 per cent in the sand immediately offshore of Coronado Strand, in the belt south of Point Loma, and in a wide strip off the Mexican portion of the area. Off the Tia Juana River calcium carbonate increases toward the outer part of the shelf until, at Loma Sea Valley, about 15 per cent is present. The sediments on the top of Coronado Bank are characterized by very high percentages of calcium carbonate, usually more than 50 per cent. A high content also exists on the shelf edge south of Coronado Canyon. Beyond the shelf edge and on the slope and floor of the San Diego Trough the calcium carbonate content has a fairly uniform value of about 25 per cent.

Along the shore the shells apparently become so broken up by wave and current action that solution is rapid. Also, near shore the calcium carbonate is much diluted by detrital sediment. Farther offshore, where detrital sediment is deposited less rapidly, as indicated by the finer grain size, calcium carbonate is more abundant. Still farther out, Loma Sea Valley and Coronado Canyon trap or divert detrital sediment coming from the mainland, so that little of it reaches the top of Coronado Bank. As a result, the sediments of the bank contain a high percentage of calcium carbonate. Other rocky areas which lack clastic deposits, such as the area directly west and south of Point Loma and the outer shelf near Los Coronados Islands, are also characterized by high calcium carbonate content. Dilution by fine sediment offshore in the San Diego Trough results in a slightly decreased percentage of calcium carbonate. The fairly uniform conditions of deposition in the trough are reflected by a uniform percentage of calcium carbonate.

ORGANIC MATTER

Organic carbon was determined for 26 samples by the empirical method of oxidation with potassium dichromate and back-titration with ferrous ammonium sulfate, as described by Allison (1935). These values were converted to total organic matter by multiplying by 1.7, a factor used by soil chemists. The highest values of total organic matter were found for four samples from the escarpment. These averaged 4.4 per cent. Six samples

from the bank top averaged 1.1 per cent, three from the bottom and sides of Loma Sea Valley averaged 2.0 per cent, twelve from the shelf inshore from Loma Sea Valley averaged 0.5 per cent, and the single sample from the beach contained no organic matter. Thus there is an increase of organic matter from shore toward deep water (fig. 10). This is in agreement with results obtained by Trask (1932) in a more general survey of the sea floor off southern California.

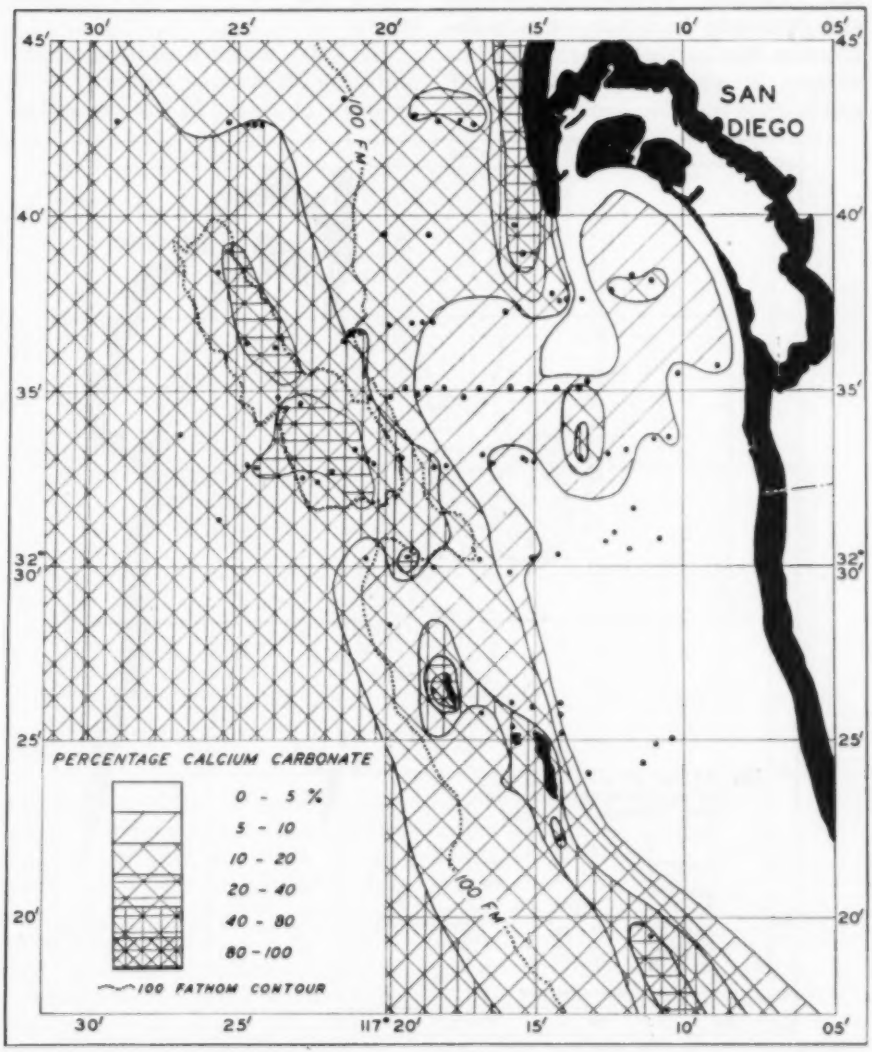


FIG. 12.—Distribution of calcium carbonate in unconsolidated sediments

MINERAL COMPOSITION

DISTRIBUTION OF TOTAL HEAVY MINERALS

Separation of heavy minerals in bromoform was made on the insoluble residues of 95 unsieved sediment samples. The results, computed on a whole-sample basis and plotted in chart form, show a distribution that is complicated by local differences in organic calcium carbonate and in grain size. Computed on a whole-sample basis, the mean value of the heavy-mineral content is 4.4 per cent, whereas on a calcium carbonate–free basis it is 6.2 per cent. The latter is a better figure for use in estimating provenance, because the calcium carbonate is not detrital but a diluent of organic origin.

The heavy minerals expressed in percentage of the insoluble residues (fig. 13) are most abundant in two zones. One

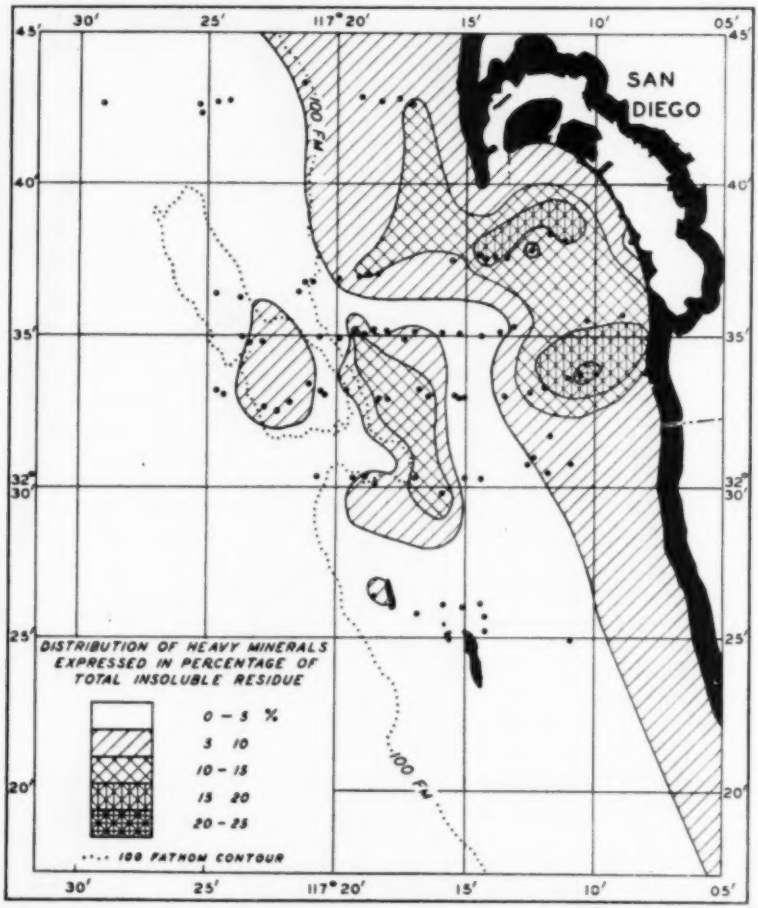


FIG. 13.—Distribution of heavy minerals in unconsolidated sediments expressed in percentage of insoluble residue.

zone (of 15–25 per cent) is near shore in the broad open bay off Coronado Strand and the Tia Juana River, and the other (of 10–15 per cent) is in deeper water skirting the head of Coronado Canyon. The highest individual percentage occurs immediately off the Tia Juana River, which is believed, therefore, to be the most important source of the heavy minerals. In the first zone the whole samples consist of fine sand, whereas in the second the samples are of very find sand to coarse silt. The coarse sand of the intervening area has less than 5 per cent of heavy minerals. Similarly, offshore on the escarpment and in the bottom of San Diego Trough the silt and clay have less than 5 per cent of heavy minerals.

A comparison of figures 8 and 13 shows that the grain size exerts very great control over the content of heavy minerals. The highest percentages of heavy minerals occur in fine to very fine sand, whereas low percentages occur in both coarse sand and fine silt and clay. This relation must mean that the heavy minerals have a narrower size range than the light minerals and that most of the heavies are restricted to the finest sand sizes. Sieving of the heavy-mineral separates showed that a median of 70 per cent by weight of the heavy minerals lay between $\frac{1}{16}$ and $\frac{1}{4}$ mm. in diameter. Figure 14 further illustrates the dependence of heavy-mineral content on grain size. It is apparent that the heavy minerals are more than twice as abundant in samples having a median diameter between $\frac{1}{16}$

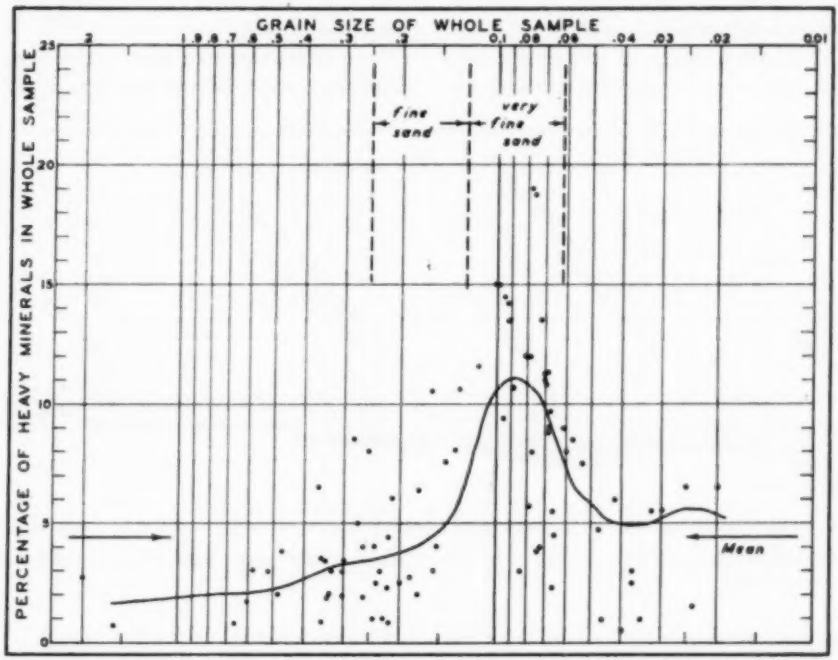


FIG. 14.—Dependence of percentage of heavy minerals of whole sample upon median diameter of whole sediment.

and $\frac{1}{4}$ mm. than in either coarser or finer sediments. In addition, the median percentage of heavy minerals in the separated $\frac{1}{16}$–$\frac{1}{4}$-mm.-size fraction of each sample was found to be 9.2, more than twice the over-all percentage in the whole sample. An even greater dependence on grain size would have been indicated, had the median diameters of the insoluble residues been available.

This relationship of percentage and composition of heavy minerals to grain size has been known for a long time, but usually it is ignored in heavy-mineral investigations. Its great importance is also shown in Van Andel's (1950) investigation of Rhine River sediments.

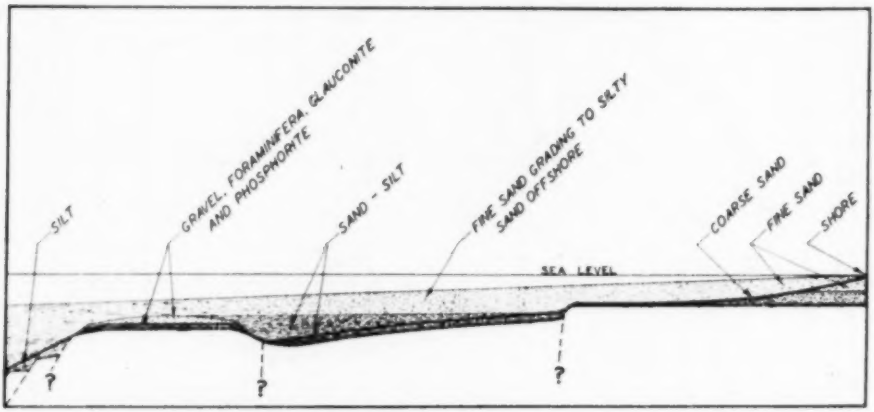


FIG. 15.—Diagrammatic cross section of present and future sea floor off San Diego. Heavy line indicates present sea floor.

MINERAL PROVINCES

General.—The minerals of 34 samples were identified in an attempt to differentiate further between the various sediment types and to investigate their possible sources. Owing to the difficulty of identifying minerals of silt and clay size, only samples from the areas of coarser sediment were studied. These samples were selected so as to conform as closely as possible to the major sediment province already discussed. The samples are grouped as follows: 4 from Coronado Strand, 9 from the medium to coarse gray- and brown-sand areas, 11 from the fine-gray-sand areas, 7 from the foraminiferal sand of Coronado Bank, and 3 from the shell sand surrounding Los Coronados Islands.

The percentages of quartz, orthoclase, plagioclase, and other constituents of the light fraction were identified chiefly by selective stains (Gabriel and Cox, 1929) and the heavy minerals by the usual optical methods. Percentages are based on counts of approximately 400 grains in each sample. Throughout the study, the whole heavy-mineral separation was used rather than a sieve fraction.

Because of their petrographic similarity, the individual analyses of samples from each province have been averaged, and the results are presented in table 4 under the headings of "Organic," "Authigenic," "Light Detrital," and "Heavy Detrital" minerals. One of the most in-

TABLE 4

MINERAL COMPOSITION OF MODERN SEDIMENTS

Mineral Species	Coronado Strand (Av. of 4 Samples) (Per Cent)		Medium to Coarse Gray and Brown Sand (Av. of 9 Samples) (Per Cent)		Fine Gray Sand (Av. of 11 Samples) (Per Cent)		Foraminiferal Sand, Coronado Bank (Av. of 7 Samples) (Per Cent)		Shell Sand Los Coronados Islands (Av. of 3 Samples) (Per Cent)	
Organic:										
Calcite	.....	2.4	.....	3.9	.....	9.0	.....	37.2	.....	36.4
Authigenic:										
Glauconite	0.0		0.0		0.0		24.0		0.0	
Clay casts	0.0		0.0		0.0		75.9		0.0	
Collophane	p*		p		p		0.1		p	
	p	p	p	p	p	p	100.0	17.4	p	p
Detrital lights:										
Quartz	49.7		56.8		70.8		69.0		48.5	
Plagioclase	38.4		35.8		25.8		26.2		30.9	
Orthoclase	11.5		7.0		2.6		4.8		18.1	
Muscovite	0.0		0.3		0.8		0.0		0.0	
Rock fragments	0.4		0.1		0.0		0.0		2.5	
	100.0	88.4	100.0	91.6	100.0	80.9	100.0	42.6	100.0	60.2
Detrital heavies:										
Hornblende	72.1		70.4		59.4		58.0		21.1	
Actinolite	5.2		4.1		8.1		4.7		2.6	
Magnetite	4.9		3.5		3.8		0.9		0.0	
Ilmenite	3.8		4.9		3.7		4.6		3.7	
Barite	2.6		3.0		1.6		0.8		0.6	
Hypersthene	2.6		3.9		2.8		0.5		0.0	
Biotite	1.8		1.8		14.2		0.6		0.0	
Epidote	1.6		1.4		1.7		4.1		10.1	
Zoisite	1.1		0.2		0.7		2.6		3.1	
Garnet	0.8		1.0		0.2		1.3		1.8	
Sphene	0.8		2.7		1.5		8.8		10.1	
Apatite	0.5		0.2		0.3		0.9		0.0	
Leucoxene	0.5		0.7		0.3		1.4		4.3	
Chlorite	0.5		0.3		0.5		1.1		1.9	
Basalt. hb.	0.4		0.4		p		0.5		0.0	
Zircon	0.4		0.2		0.4		0.4		0.0	
Glaucophane	0.2		0.0		0.3		0.5		0.0	
Olivine	0.2		0.7		0.1		1.9		0.0	
Enstatite	0.0		0.4		p		0.0		0.0	
Hematite	0.0		0.2		0.1		0.0		2.0	
Limonite	0.0		p		0.0		0.0		0.6	
Anatase	0.0		0.0		0.3		0.0		0.0	
Alteration	0.0		0.0		0.0		5.7		36.3	
Augite	0.0		0.0		0.0		0.5		1.8	
Diopside	0.0		0.0		0.0		0.3		0.0	
	100.0	9.2	100.0	4.5	100.0	10.1	100.0	2.8	100.0	3.4
		100.0		100.0		100.0		100.0		100.0

* "p" indicates present in amounts less than 0.1 per cent.

teresting general features shown by these data is the high percentage of feldspar and the marked abundance of plagioclase relative to orthoclase. Plagioclase is about three times as abundant as orthoclase, and the total feldspar content of all samples averages about 30 per cent; thus the sediment is properly termed an "arkose." The great proportion of feldspar in itself is not unusual, since most of the Cretaceous and Cenozoic sandstones in California are feldspathic (Reed, 1933, pp. 276–277), owing to the still higher feldspar content of the batholithic source rocks. However, it is of interest that sediments very similar to those of past geologic periods in California are present on the sea floor today.

Inshore areas.—The mineral assemblages of the inshore areas (Coronado Strand, the medium to coarse gray and brown sand, and the find gray sand) are very similar (table 4) and therefore can be discussed together. Unlike the Coronado Bank and Los Coronados Islands areas, the sediments of the inshore areas are almost entirely detrital. The most striking feature of their heavy constituents is the abundance of green hornblende. This, together with the high percentages of actinolite, magnetite, barite, and hypersthene and the low percentages of epidote, zoisite, and sphene, suggests that the sands of the inshore areas have a common source; but this does not necessarily mean that they are of the same age.

The only notable exception to the general similarity of minerals in all the inshore areas is the unusual abundance (14.2 per cent) of brown biotite in the fine gray sand southeast of Point Loma. Because of its platy habit, biotite is easily removed from the beach by wave and current action and carried to deeper water. The fact that biotite is more easily transported than most other minerals is apparent from the large size of the biotite flakes in the fine gray sand. Most of these flakes are several times larger than the median diameters of the whole samples. The same explanation applies to the absence of appreciable amounts of biotite in the coarser-sand zones off the mouth of the Tia Juana River, where the presence of medium to coarse sand may indicate that the currents in that region are capable of keeping biotite in suspension with the finer materials until they reach deeper water.

Coronado Bank.—In contrast to the predominantly detrital sediments of the inshore areas, the sediments flanking Los Coronados Islands and those lying on top of Coronado Bank are distinguished by their high percentages of organic constituents. As previously mentioned, the sediments in the vicinity of the islands consist largely of coarse shell fragments, and those on the bank are chiefly foraminiferal tests. Although the organic calcium carbonate content of many samples from these areas exceeds 50 per cent, the average percentage of calcareous material in the samples selected for mineral analysis is 37.2 per cent for the Coronado Bank samples and 36.4 per cent for the Los Coronados Island area samples.

In addition to its high percentage of calcareous organic sediment, Coronado Bank is further distinguished by an appreciable quantity of authigenic minerals. Chief of the latter are internal casts of clay and glauconite in many of the foraminiferal tests. These casts range from light buff through tints of green to dark green and black. This color transition suggests that there may be a genetic connection between the clay casts and the formation of glauconite, but proof of this relation must await chemical and X-ray studies. Some of the glauconite occurs in the form of coprolites, but, be-

cause of their scarcity, they have not warranted a separate entry in table 4. Another minor authigenic constituent is collophane, which probably has the same origin as the larger phosphorite nodules of the bank. It seems probable that the formation of this mineral is confined to the bank and that the few grains of collophane in samples from the other areas are detrital.

The detrital constituents of the Coronado Bank sediments are considerably more rounded and more highly weathered than those of the inshore provinces. It is inconceivable that the coarse bank sediments could be transported across Loma Sea Valley and deposited on top of Coronado Bank under present conditions. The same general suite of heavy minerals characteristic of the inshore areas is also present in the bank sediments, but there are some notable differences in their percentages. Highly altered grains, expressed as "alteration" in table 4, are fairly common (5.7 per cent of the heavies) in the bank samples, but no grains of this type were found in samples from the inshore areas. Similarly, sphene, epidote, and zoisite are considerably more abundant on Coronado Bank, and magnetite and hypersthene, which are common in the inshore areas, are rare on Coronado Bank. These factors suggest that the bank sediments have a different source and are of greater age than some of the inshore sediments.

Los Coronados Islands area.—The detrital minerals of the shell sand surrounding Los Coronados Islands, like those of Coronado Bank, exhibit a greater degree of rounding and are more highly weathered than the minerals of the inshore provinces. The general coarsening of these sediments toward the islands suggests that they are locally derived. The most striking difference between the heavy minerals of this province and those of the other mineral areas is the comparatively low percentage of green hornblende and the great abundance of altered minerals (compare tables 1 and 4). Titanite, epidote, and zoisite are much more prominent in the island area, whereas magnetite and hypersthene, which are common in the inshore areas, are rare or absent. The occurrence of a considerable amount (4.3 per cent) of leucoxene and the marked decrease in actinolite further differentiate the Los Coronados Islands area sediments from those inshore.

Conclusions

Topography and structure

Knowledge of the structure of an area depends on detailed geological mapping, which is impracticable beneath the sea at the present time. Accordingly, only the larger structural framework can be discerned, chiefly from the topography. On the adjacent land the sedimentary rocks overlap the basement complex to the east and dip prevailingly to the west at low angles. Folding is minor, but faults complicate the structure, at least north of San Diego.

The dominant topographic features of the submarine area are: (1) the narrow shelf; (2) the elongate ridge rising as an island group on the outer edge of the shelf and progressively deepening to the north as a shallow ridge, a flat-topped bank, and a narrow deep ridge; (3) the longitudinal valley and the transverse submarine canyon which separate the bank and deep ridge from the shelf; (4) the steep straight escarpment bordering the shelf and the bank; and (5) the broad flat trough at the base of the escarpment. The latter resembles escarpments on the

flanks of mountain ranges in southern California that have usually been interpreted as fault scarps, and it is also similar to others of the sea floor that give the region the appearance of a submerged basin and range province wherein faulting predominates over folding.

Coronado Bank is bounded on the seaward side by Coronado Escarpment and on the landward side by Loma Sea Valley. This valley is relatively straight, it parallels the escarpment, and it is asymmetrical in cross section. Extensive faulting is present on Los Coronados Islands nearly on its strike. These features suggest that the valley was cut along a fault or was formed by faulting, either case leading to a separation of the bank from the shelf. The fact that the top of the bank is locally somewhat shallower than the edge of the shelf suggests that it may have been upthrust between two bounding faults. It may be part of a sliver block at the edge of the shelf, whose uptilted end is marked by Los Coronados Islands and whose downtilted end is the deep ridge north of the bank. The age of deformation in the submarine section cannot be determined with any great degree of accuracy, though some of it must have been post-Pliocene, because late Pliocene to late Pleistocene strata have been affected. Most of it must have been pre-late Pleistocene, however, because of the general similarity of depths along the shelf and on the bank, that surface being at least as old as late Pleistocene.

Coronado Canyon, that completes the isolation of Coronado Bank, is transverse to the structural trend, is winding in plan, and has a relatively even outward slope. If the canyon and valley system is erosional, it is likely that Loma Sea Valley was formed first along a fault zone and that Coronado Canyon cut across the ridge at a low point and captured the stream flowing down the valley. Conceivably, such piracy may have resulted from turbidity currents, but the topography is more suggestive of stream erosion, because the ridge would have been a poor source of sediment for turbidity currents. A delta outside the water gap is suggested by the deep and flat top of the great fan which extends out into San Diego Trough at the mouth of Coronado Canyon. If the area were once about 500 fathoms higher than now, stream erosion could have produced the valley system and submarine landslides, and/or turbidity currents could have modified the delta and kept the canyon open after submergence.

AGE OF BEDROCK

Bedrock was obtained from three main areas of the sea floor off San Diego: Coronado Bank, the shelf near Los Coronados Islands, and an area just south of Point Loma. The rocks of the top of Coronado Bank contain Foraminifera and mollusks of late Pliocene to Recent age and have the same suite of heavy minerals as do the Pliocene formations ashore. The beds are truncated by an erosional surface that probably was cut during the last glacial stage of lowered sea-level, thus the rocks are probably older than late Pleistocene. Rocks from the shelf near Los Coronados Islands are unfossiliferous but are lithologically similar to the Miocene rocks of the islands. In the area south of Point Loma the rocks, predominantly shales, contain Cretaceous Chico Foraminifera and are lithologically similar to Cretaceous rocks that crop out on land at the tip of Point Loma.

AGE AND SOURCE OF GRAVELS

Gravels that range from pebbles to boulders are present in all areas of bed-

rock. The fragments are of two types: sandstone and shale, like the underlying bedrock, and igneous and metamorphic rocks that are partly from underlying conglomerates and partly derived from land and mixed with those of local origin. The gravels are believed to have been deposited during the last stage of glacial lowering of sea-level, because (1) they lie atop a bedrock surface cut probably during the last glacial epoch, (2) some pieces are much too large to be transported or eroded from bedrock by the weak bottom currents of today, (3) associated with them are a few shells that are characteristic of colder water than now covers the area, and (4) they contain abundant pholad borings that are probably characteristic of shallower depths than exist today.

In addition to gravels on bedrock, there are several areas of gravel where no bedrock is known. Largest of these is the boulder patch off the mouth of the Tia Juana River and an area of scattered cobbles off Coronado Strand. These areas are believed to be part of the general distribution during the late Pleistocene. A blanket of gravel may have been deposited throughout the area at that time, but it has been buried beneath finer sediments except in these few areas.

SOURCES OF SEDIMENT

Although the finer unconsolidated sediments have been described in some detail, the investigation is not complete without some inquiry into the sources of the detrital part of the sediments. To aid in this study, all available data on the mineral constituents of the rocks from which the sediments may have been derived are summarized in table 1 for comparison with table 4.

There is a general similarity in mineral composition of all the formations from Cretaceous to Recent, including the unconsolidated sediments of the offshore area. Differences in the percentages of these mineral constituents serve to differentiate the formations, but the general similarity of mineral species strongly suggests that all were ultimately derived from a common source. High percentages of feldspar, hornblende, biotite, and other igneous accessory minerals indicate that the formations have been derived chiefly from intermediate igneous rocks, probably the tonolite batholith that forms the core of the Peninsular Range on the east. Supporting evidence for a relatively local origin is also found in the coarseness, the angularity, and the general freshness of the grains that make up the Cretaceous and Cenozoic sedimentary rocks.

A comparison of the mineral data listed in tables 1 and 4 shows a close resemblance between the medium to coarse gray and brown sand, the fine gray sand, and the beach sands of Coronado Strand and of the whole coast from Point Descanso to Oceanside. As some of the inner-shelf sediments are probably derived from the beaches, it is not surprising that they should correspond so closely.

There is also a close correspondence between the unconsolidated sediments of the inner-shelf areas and the marine Pliocene formations on the mainland. The general similarity in abundance of green hornblende and in percentage of actinolite, biotite, epidote, and ilmenite is striking. It seems quite probable that a considerable proportion of the inner-shelf sediments have been derived from the mainland Pliocene formations because of their wide distribution and their susceptibility to erosion. However, it is equally apparent that some of the inner-shelf sediments must have been contrib-

uted by another source, possibly the igneous and metamorphic rocks exposed in the Tia Juana River drainage basin.

In many respects the detrital constituents of the Coronado Bank province appear to be even more closely related to the Pliocene formations on land than are the inner-shelf sands. Such diagnostic minerals as actinolite, sphene, ilmenite, epidote, and zoisite of the Pliocene formations are present in about the same proportions. There are, however, some significant differences in a few constituents. Hornblende, although it is the most prominent heavy mineral in the Pliocene formations, is more abundant in the bank sediments, and biotite, which is fairly common in the Pliocene, makes up only a very small percentage of the bank sediments. Possibly the biotite has been winnowed out of the bank material by current action, causing an increase in the percentage of hornblende. Evidently the detrital sediments on top of Coronado Bank are either reworked Pliocene or were derived from the same general source as the Pliocene formations. Reworking from the underlying Pliocene rocks is considered more probable because of the difficulty of transportation to the bank after its separation from the shelf.

The sediments surrounding Los Coronados Islands have high percentages of epidote, sphene, and altered minerals. These minerals suggest derivation either from Eocene formations on land or from the Miocene rocks of the islands and vicinity. Owing to their coarseness, their high content of rock fragments, and their proximity to the islands, it is more probable that they came from the islands. The presence of hornblende, although much less abundant than in the other mineral areas, probably indicates some admixture of the sediments from the shelf.

CONDITIONS OF DEPOSITION OF THE INNER-SHELF SEDIMENTS

Several lines of evidence suggest that at least two periods of deposition are represented in the finer materials that cover the inner shelf. The first and oldest of these periods is represented by the medium to coarse brown sand and possibly the medium to coarse gray sand, because (1) it is highly improbable that present-day waves and currents in the area can transport coarse sands so far from shore (more than 7 miles) and especially to depths exceeding 20 fathoms; (2) the brown sand is deeply stained with ferric iron oxide and is more highly weathered than other sediments of the inner shelf; (3) fine gray sand, which is now supplied by the Tia Juana River, lies shoreward of the brown-sand area and probably overlaps it.

The general north-south trend of the brown-sand area and its extension in isolated patches to the tip of Point Loma suggest that it was deposited when the shoreline was considerably west of the present position at a time of relatively lower sea-level. The close association of the brown sand with rounded cobbles and boulders very similar to the conglomerates and deeply iron-stained sands covering many of the terraces of the mainland suggests further that the coarse sands on the inner shelf are reworked terrace deposits. According to Ellis and Lee (1921) and Hanna (1926), the prominent stream valleys bordering the ocean were formerly cut to depths of 100–200 feet below their present levels. Alluvial fill in the valleys consists of variable sand and silt atop a layer of well-rounded boulders and cobbles that varies in thickness from 10 to 35 feet. The marked similarity between the boulders, cobbles, and coarse sand at the bottom of the Tia Juana River Valley and the sediments of the

coarse sand and the boulder zones in the inshore area is striking.

One hypothesis worthy of consideration is that, when the sea was lowered to the level of the outer shelf and Coronado Bank during a late Pleistocene glacial epoch, extensive nonmarine terrace deposits accumulated on the exposed parts of the shelf. Later, as the sea-level rose, the shore transgressed these deposits, reworking them and leaving an accumulation of coarse, brown, iron-stained sands and gravels. Probably these coarse sediments were widely distributed over most of the inner-shelf region, but only the remnants that have not been covered by present sediments are still exposed.

The second period of sedimentation in the inner-shelf area is presented by the fine gray sand which is contributed by the Tia Juana River at the present time. Part of this detritus has been swept northward by longshore currents to form Coronado Strand, but much of it, particularly the fine sand, is being deposited on the inner part of the shelf. The very fine sand, silt, and clay can be more easily transported than the coarser sediments, so that they reach and are deposited on much of the outer part of the shelf and in the deeper parts of the offshore area.

GEOLOGICAL SIGNIFICANCE

The unusual density of sampling and the relative completeness of background information in this region make it probably the best-known submarine area of its size in the world. As such, it provides information that can be used to help interpret certain sedimentary environments that may be encountered in ancient rocks exposed in outcrops or penetrated by the drill.

The outstanding characteristic is the presence of an unconformity that is only partly buried beneath a blanket of sediment. Where still exposed, the bedrock is associated with a gravel that forms the first deposit above the unconformity. In the depressions of the irregular bedrock surface and especially near the shore, the bedrock and its cover of gravel have been buried by finer sediments. The first layer was of coarse sand; the second (now being deposited) is of fine sand. The second layer and possibly also the first layer are generally thickest in the east near their source. In areas of shallow quiet water, as in San Diego Bay, or in deep quiet water, as in Loma Sea Valley, Coronado Escarpment, and San Diego Trough, silty sediments carried in suspension were and are being deposited atop bedrock, gravel, and sand. In some localities where currents are strong, the shelf is covered by shell fragments, and the bank by Foraminifera, glauconite, and phosphorite. Only if further prograding of the fine sand and thickening of the silt forms a level depositional surface will sediments form layers that are uniform or broadly gradational. This state probably will not be reached until the whole San Diego Trough is filled.

Only the silt of the deeper water contains even moderately high organic matter that may later be converted to oil. The less rich sands interbedded with the silt may eventually serve as reservoir sandstones retaining some of the oil from the silts.

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PHASE RELATIONS OF THE ALKALI FELDSPARS. II . THE STABLE AND PSEUDO-STABLE PHASE RELATIONS IN THE ALKALI FELDSPAR SYSTEM[1]

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ABSTRACT

In the first part of this paper (1952, pp. 436–450), the structural states of the pure compounds $KAlSi_3O_8$ and $NaAlSi_3O_8$ have been discussed. This part considers the mixed-crystal series of these compounds and their unmixed states.

This section will be simplified in the following way. Adularia and "common orthoclase" will be disregarded at first, inasmuch as they are considered unstable phases.[2] In like fashion, any distinction between sanidine (low) and sanidine (high) will be neglected, as it does not influence the principle to be proposed (see the first part of this paper, Laves, 1952, pp. 446–449). The phases thus to be considered are (1) sanidine, $KAlSi_3O_8$, disordered *re* Al, Si; (2) analbite, $NaAlSi_3O_8$, disordered *re* Al, Si; (3) microcline, $KAlSi_3O_8$, ordered *re* Al, Si; and (4) albite, $NaAlSi_3O_8$, ordered *re* Al, Si.

PHASE RELATIONS IN THE SERIES SANIDINE-ANALBITE

RELATIONS UNDER CONDITIONS OF RAPID COOLING

Schairer and Bowen (1935) showed that a homogeneous mix-crystal series of alkali feldspars was preserved at room temperature when the originally molten material was crystallized and cooled. A complete series of solid solutions from $KAlSi_3O_8$ to $NaAlSi_3O_8$ appeared evident. The recent and more extensive work of Bowen and Tuttle (1950) confirmed this complete series by the use of X-ray powder data. The lattice constants of $KAlSi_3O_8$ were found to differ measurably from those of $NaAlSi_3O_8$ (especially the *a*-axis; see Taylor, 1933; Taylor, Darbyshire, and Strunz, 1934), and, using Vegard's law to determine the composition of the crystalline products, the Bragg angle of $(\bar{2}01)$ was measured. Bowen and Tuttle showed that the *d*-spacing of $(\bar{2}01)$ plots as a straight line as a function of composition. As no indication of a discontinuity in the curve was observed, they concluded that the solid-solution series was complete, with no suggestion (experimentally) of a symmetry change. From these data it again appears that the change from the monoclinic symmetry of $KAlSi_3O_8$ to the triclinic symmetry of $NaAlSi_3O_8$ takes place without a break in any observed properties.

The puzzling nature of this series had been used earlier as evidence in favor of Mallard's hypothesis, i.e., that the monoclinic symmetry was only apparent rather than real and was due to submicroscopic twinning. It thus appeared desirable to the writer to obtain more conclusive evidence as to the crystallographic behavior of this series. Consider-

[1] Manuscript received May 31, 1952.

[2] The terms "stable," "metastable," and "unstable" have their usual thermodynamic meaning. A phase is, in this paper, called "pseudostable" if it evidently should be considered unstable but its velocity of change to the stable state is too slow to be observed in the course of the experiments.

ation of the linear change of the $(\bar{2}01)$ spacings with composition shows that this criterion is not valid as an indication of the presence or absence of a monoclinic $\rightleftharpoons$ triclinic symmetry change. This is because the spacing of $(\bar{2}01)$ depends only upon the a- and c-axes and the angle β. Therefore, any change in the angles α and γ (precisely, those angles that deviate from 90° in a monoclinic-triclinic change) would not be expressed in the $(\bar{2}01)$ spacing. The fact that analbite deviates rather strongly from monoclinic symmetry suggested to the writer that some break in the series should be present and should be observable.

In the present investigation, only natural feldspars were used, all the X-ray work being single-crystal diffraction. For the information desired it did not seem necessary to go through the difficult work involved in obtaining a series of synthetic materials (see Franco and Schairer, 1951, p. 261). Natural crystals, however, always have some anorthite in solid solution, a fact that had to be taken into consideration.

TABLE 1

COMPOSITION OF THE FELDSPARS HERE INVESTIGATED

No.*	Or (Per Cent)	Ab (Per Cent)	An (Per Cent)	Method of Analysis†	"OrAb" Content
2	17.9	79.9	2.2	Spectroscopic	$Or_{19}Ab_{81}$
	19	81		X-ray	
3	23.8	71.8	4.4	Spectroscopic	$Or_{23}Ab_{77}$
	23	77		X-ray	
4	35	65		X-ray	$Or_{35}Ab_{65}$
5	37	63		X-ray	$Or_{37}Ab_{63}$

* Nos. 2 and 3 were kindly supplied by T. F. W. Barth, who collected them as crystals (~1-cm. size) presumably ejected from a volcano at "Army Lookout at First Bluff, St. George–Pribilof Island." Nos. 4 and 5 were taken from a rock kindly supplied by N. L. Bowen, who described it (1937, p. 15) as phonolitic trachyte from Athi Plains, near Nairobi, East Africa (N.L.B. 303).

† The spectrographic analyses were made by Oiva Joensuu, the limit of error being 10 per cent of the CaO present. The X-ray analyses were made by the writer, using Vegard's law on the (400) reflection in single-crystal photographs. For more details see legend of fig. 3. Sample numbers are the same as in table 2.

It has long been known that the intermediate mixes of Na-K feldspar crystals (between the approximate compositions $Or_{80}Ab_{20}$-Or_5Ab_{95}) are not stable at low temperatures. It is now well known that in this range the feldspars normally formed in nature are perthitic and that those that appear homogeneous are actually cryptoperthitic, as first suggested by Brögger (1882, pp. 262, 292). These crystals, therefore, cannot be used, as they are inhomogeneous and are not representative of the high-temperature solid-solution series. Fortunately, however, heat treatment of these unmixed crystals can produce homogenization (see Spencer, 1937) and the resulting "sanidines," when cooled quickly, remain homogeneous.

At first sight it appeared simple to determine α and γ on homogenized natural crystals, these two lattice constants being the important ones for the evaluation of symmetry with changing Na/K ratio. It soon became obvious, however, that the Ca content of the Na-rich crystals was generally high enough that they could not be considered representative of the pure alkali feldspar series. However, five compositions in the series, rather low in Ca, were obtained and proved useful for this investigation.[3] The compositions of these crystals are shown in table 1. For the present purpose, the anorthite content is included with the albite.

Single-crystal X-ray photographs (Buerger precession technique) showed that crystals 2 and 3 were homogeneous, whereas 4 and 5 were not. The inhomoge-

[3] Materials 2 and 3 were kindly supplied by T. F. W. Barth and materials 4 and 5 by N. L. Bowen.

TABLE 2

LATTICE CONSTANTS*

No.	Or×100/(Or+Ab+An)	α^* (001)/(010)	β^* (100)/(001)	γ^* (100)/(010)	$d(100)=1/a^*$	$d(010)=1/b^*$	$d(001)=1/c^*$	α	β	γ	a	b	c	Sample
	High-Temperature Modification													
1	0	86° 02′	63° 27′	88° 15′	7.296	12.823	6.364	93° 34′	116° 30′	89° 59′	8.115	12.871	7.127	Analbite
2	19	87 04	63 35	88 31	7.379	12.919	6.395	92 32	116 23	90 12	8.239	12.940	7.147	Analbite
3	23	87 22	63 32	88 40	7.395	12.936	6.402	92 17	116 26	90 11	8.260	12.950	7.157	Analbite
4	35	88 38	63 40	89 18	7.444	12.974	6.421	91 11	116 21	90 06	8.308	12.981	7.166	Analbite
5	37	89 15	63 42	89 35	7.454	12.976	6.421	90 38	116 19	90 06	8.316	12.977	7.163	Analbite
6	92	90	64 00	90	7.697	13.030	6.449	90	116 00	90	8.564	13.030	7.175	Orthoclase, heat-treated (Cole *et al.*, 1949)
	Low-Temperature Modification													
7	92°	90°	63° 59′	90°	7.695	12.996	6.464	90°	116° 01′	90°	8.562	12.996	7.193	Orthoclase, original (Cole *et al.*, 1949)
8	89	90	63 55	90	7.696	12.964	6.482	90	116 05	90	8.568	12.964	7.216	"Adularia"
9	90	90° 25′	64 04	92° 16′	7.707	12.956	6.496	90° 39′	115 56	87° 42′	8.577	12.967	7.223	Microcline
10	0	86 22	63 26	90 30	7.299	12.736	6.396	94 19	116 40	87 37	8.156	12.773	7.170	Albite (Amelia)
11	0	86 25	63 34	90 27	7.279	12.753	6.389	94 14	116 31	87 43	8.135	12.788	7.154	Albite (Kodarma) (Cole *et al.*, 1951)

* The limit of error for Nos. 1–5 and 8–10 is less than 5′ for angular measurements and less than 0.2 per cent for spacings. The relative values are more accurate. Values in Å.

neous crystals were homogenized readily in approximately 15 minutes by heating at 800°–1,000° C. The lattice constants determined on crystals 2–5 are compiled in table 2. These values are plotted in figures 1–4, along with the constants of pure analbite, some of which were published previously (1950). If one plots those lattice constants that are sensitive to a monoclinic-triclinic symmetry

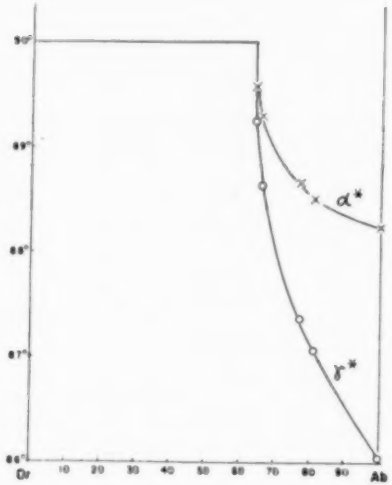

FIG. 1.—Angles $a^* = (010)/(001)$ and $\gamma^* = (010)/(100)$ as functions of composition for crystals 1–5 of tables 1 and 2.

change, the discontinuous change of symmetry (at room temperature) at approximately $Or_{63}Ab_{37}$ is shown strikingly (fig. 1). The trend of these curves indicates a second- (or higher-)order transformation, as defined by Ehrenfest (1923). It should be noted, however, that only the composition is used as a variable here, temperature remaining constant.

Microscopic observation of the twinning of crystals 2 and 3 indicated to the writer that these crystals had gone through a monoclinic-triclinic inversion. The crystals, which were ejected directly from a volcano, were obtained from T. F. W. Barth.

The optical appearance of samples 2 and 3 (even before heat treatment) showed puzzling features: (1) Portions of the crystals were untwinned but had an observable extinction angle. (2) Other portions showed fine albite twinning in places, pericline twinning in other places, as well as combination albite and pericline twinning. The fine cross-hatching of the combination twins made the crystals look like microcline. (3) In some places the fine twinning graded into an area

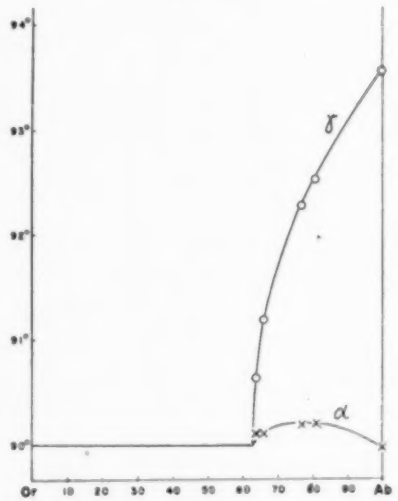

FIG. 2.—Angles α and γ as functions of composition for crystals 1–5 of tables 1 and 2.

where no twinning was visible, and these areas showed virtually parallel extinction.

X-ray photographs were taken of all the areas described above. The untwinned areas showing inclined extinction (1) produced a single set of diffraction spots. The areas of (2) produced *two* or *four* sets of spots corresponding to either or to the combination of twin laws. The

areas of (3) gave results similar to those of (2), but the spots were somewhat diffuse in some cases.

These observations can be interpreted in the way used by the writer (1950) for microcline. It was deduced in this case

In addition, the following phenomena were observed: (4) Heat treatment produced changes in the twinning of the crystals. This was observed both optically and on X-ray photographs. Several minutes at 800° C. were sufficient to

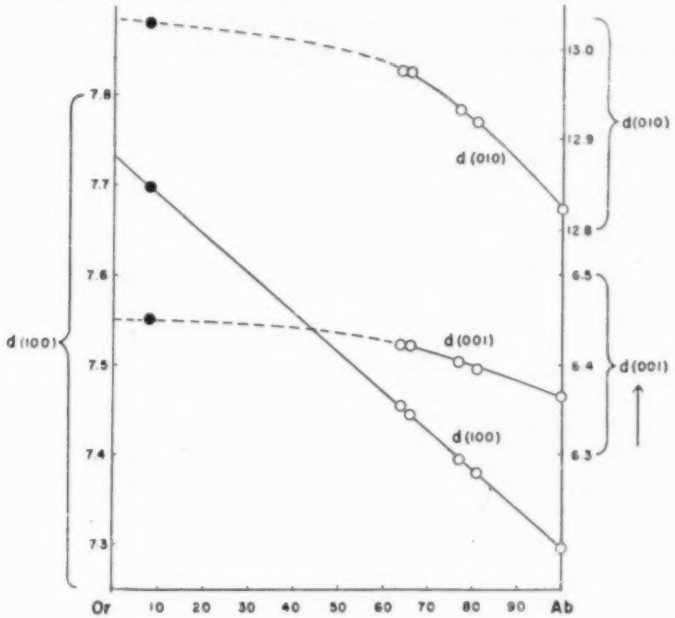

FIG. 3.—Spacings of (100), (010), and (001) as functions of composition for crystals 1–5 of tables 1 and 2. The solid circles are plotted from data of Cole, Sörum, and Kennard (1949); see no. 6 in table 2. The straight line drawn for $d(100)$ is used as a determinative curve for composition [in analogy to the use by Bowen and Tuttle (1950) of (201); they showed that on synthetic material, $d(201)$ as a function of composition follows Vegard's law very well]. The possible errors with respect to an accurate determination of composition caused by this procedure are considered to be negligible for the purpose of this investigation. The results of composition determinations by this procedure are in satisfactory agreement with spectrographic results given in table 1. Values in Å.

that the twin combination of albite and pericline laws is a consequence of an inversion from monoclinic to triclinic symmetry *if* the twin axis of the pericline twins is perpendicular to the twin plane of the albite twins. Crystals 2 and 3 showed these same relations, and it was concluded, therefore, that they had once been monoclinic.

change the whole twin pattern: albite twinning could go over to pericline twinning, and, vice versa, untwinned areas could become twinned, etc. (5) The extinction angles [against (010)] of untwinned or broadly twinned portions of these crystals were measured as a function of temperature. A polarizing microscope equipped with a heating stage was

used.[4] The homogeneous triclinic crystals were observed to become monoclinic upon heating. The extinction angle changed appreciably with increasing temperature, as seen in figures 5 and 6. The temperature at which the change from triclinic to monoclinic symmetry took place could be determined with an accuracy of $\pm 8°$ C. (crystals 2 and 3) and $\pm 3°$ C. (crystals 4 and 5). This symmetry change is readily reversible, and the relation of extinction angle to temperature is constant for a given specimen. (6) The high-temperature monoclinic symmetry could not be quenched in.

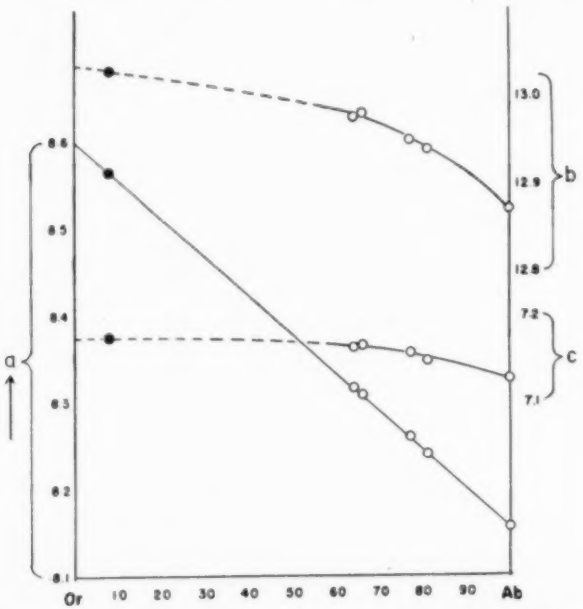

FIG. 4.—The lengths of the *a*-, *b*-, and *c*-axes as functions of composition for crystals 1–5 of tables 1 and 2. See remarks in the legend of fig. 3.

These experiments showed that, in at least a certain range of composition, alkali feldspars are monoclinic at high temperatures and triclinic at low temperatures. It is apparent that this transition is of the displacive type, such as in high $\rightleftharpoons$ low quartz or high $\rightleftharpoons$ low cristobalite, etc. If one follows general usage in naming forms having a displacive transformation, difficulties arise. The monoclinic high-temperature modification of $(K,Na)AlSi_3O_8$ is called "sanidine." These two modifications, therefore, might be called "high-sanidine" and "low-sanidine." Aside from an objection that will be discussed later, Tuttle and Bowen (1951) have already used this terminology for states that do not correspond to these.

As was mentioned, the structural geometry of these crystals at room temperature is similar to that of the high-temperature modification of pure $NaAlSi_3O_8$, here called "analbite." The

[4] The high-temperature stage was a modified version of the design published by Elizabeth A. Wood (1951).

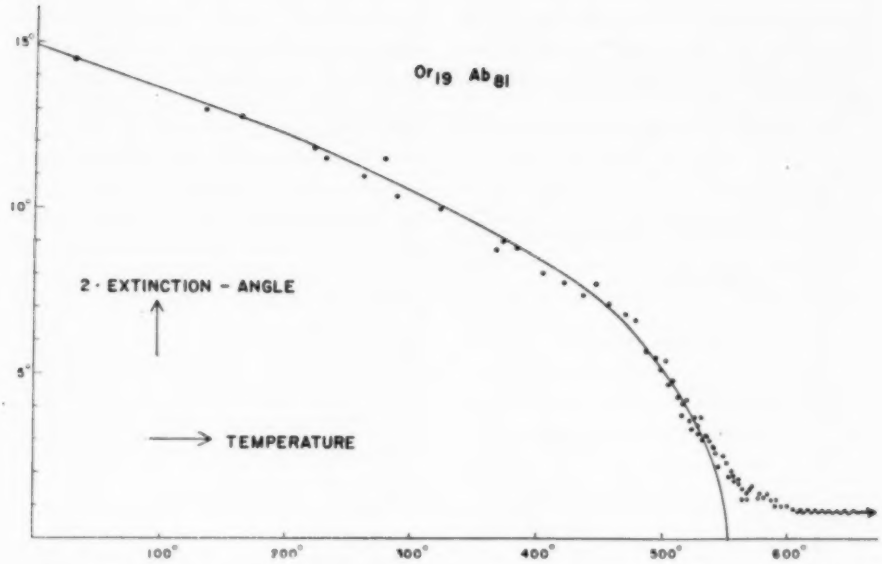

FIG. 5.—Extinction angles as a function of temperature. The measurements were made on a portion of crystal 2, tables 1 and 2. The specimen was a cleavage flake parallel to (001) and consisted of two parts in albite twin position. The plotted angles are therefore doubled extinction angles. That the values did not become zero above the extrapolated transformation temperature seemed to be due to somewhat mosaic structure, indicated by an undulous character of the extinction position at elevated temperatures.

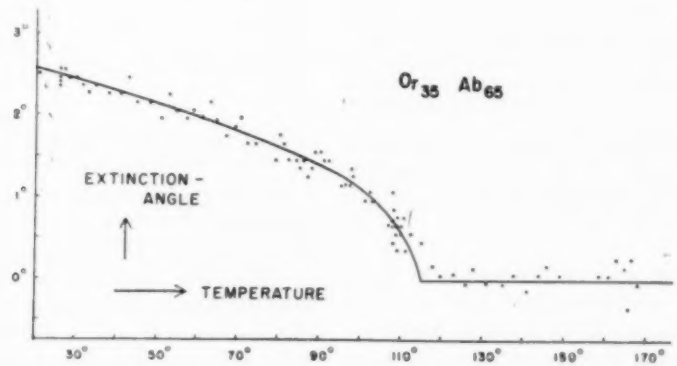

FIG. 6.—Extinction angles as a function of temperature. Untwinned cleavage flake of crystal 4, table 2, parallel (001).

pertinent crystallographic values are included in table 2. The angles α, γ, α^*, and γ^*, which are the essential values for the monoclinic-triclinic distinction, are plotted in figures 1–4.

The inversion temperatures as a function of composition are plotted in figure 7. Very similar results have been observed independently by W. S. MacKenzie (personal communication). The lines represent extrapolations to higher temperatures that are of special interest. As shown in figure 8, the inversion curve is drawn to meet the solidus curve at approximately Or_5Ab_{95}. If it were extrapolated as a straight line, the curve would hit the pure albite composition at approximately 1,080°. This would indicate that the disputed "barbierite," or monoclinic modification of $NaAlSi_3O_8$, could exist and that there would be a truly complete high-temperature series of monoclinic solid solutions between $KaAlSi_3O_8$ and $NaAlSi_3O_8$. The writer has no evidence that inversion takes place in pure analbite and follows, therefore, the conclusions drawn by Bowen and Tuttle (1950, p. 577) as far as the question of the existence of a monoclinic modification is concerned. Synthetic crystals[5] heated to 1,110° C. and examined before and after heating showed no change in twin distribution of the sort described above with respect to the natural volcanic crystals. The inversion curve was thus extrapolated to intersect the solidus, indicating that pure analbite remains triclinic. The extrapolation was made to indicate a maximum Or content

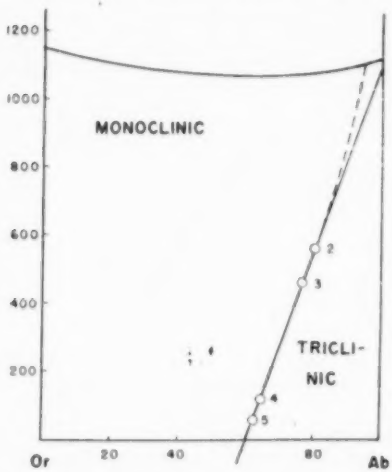

FIG. 7.—The transformation temperatures, determined as shown in figs. 5 and 6, plotted as functions of composition for the crystals 2–5, table 2. A straight line drawn through these points would meet the Ab-axis below the melting point of pure $NaAlSi_3O_8$, indicating the existence of a monoclinic modification at temperatures close to the melting point. Following Tuttle and Bowen's (1950, p. 577) discussion on the probable nonexistence of a monoclinic high-temperature $NaAlSi_3O_8$ modification (the much disputed barbierite) and without experimental evidence for its existence, the writer extrapolated the transformation curve so as to be in accordance with the opinions expressed in the recent literature. Thus the trend of the dashed line will be used for the following considerations.

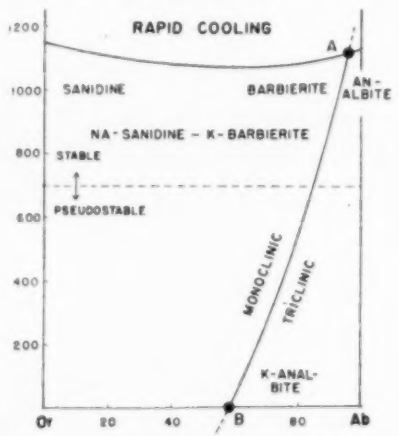

FIG. 8.—Phase relations of material cooled rapidly enough to suppress exsolution and ordering of the Al and Si ions. Points *A* and *B* are to be considered later in connection with figs. 17 and 18.

[5] Crystals prepared by I. Friedman (1951) and described by the writer in a paper on "high-albite" (1950).

for the crystals that remain triclinic below the solidus.

It is of considerable interest to note that these conclusions could have been reached (at least qualitatively) many years ago. In 1884 Förstner heated feldspars from Pantelleria on the microscope stage and observed that the Na-rich alkali feldspars changed from optically triclinic to optically monoclinic symmetry. He reported that this transformation was reversible and that the monoclinic state could not be quenched. His data show scatter, probably owing to the fact that his feldspars were partially unmixed and would produce apparent inconsistent transformation temperatures (see p. 567). Apparently no one ever checked his observations, and they have been neglected in the interpretations of the alkali feldspar relations. Although Mäkinen (1917) was aware of these data, he rejected them for use in his phase diagram because they did not fit his interpretation. He states (p. 163):

> Bei der Aufstellung des Diagramms wurden die von Förstner (Z. Krist. 9 (1884), s. 333) angestellten Experimente mit den Pantelleria-Feldspäten nicht berücksichtigt. Seine Resultate sind gar nicht mit den Tatsachen vereinbar, auf welche dieses Diagramm gegründet ist.

RELATIONS UNDER CONDITIONS OF SLOW COOLING (PSEUDO-STABLE EQUILIBRIUM)

The existence of natural perthites and experiments in which these materials were homogenized (Kôzu and Endô, 1921; Dittler and Köhler, 1925; Spencer, 1937) were more than indicative that the alkali feldspars do not form a complete solid-solution series at room temperature. Several authors early concluded that a high-temperature solid solution should unmix on cooling. The unmixing curve varied as a function of temperature and composition in different writers' interpretations, experimental evidence being lacking. Oftedahl's interpretation (1948) (see fig. 4, first part of paper, p. 437) comes closest to the excellent experimental diagram of unmixing determined by Bowen and Tuttle (1950) (see fig. 5, first part of paper, p. 437).

Using only that portion of the diagram experimentally determined by Bowen and Tuttle, the writer proposes figure 9, which is a combination stable and "pseudo-stable" diagram. The sanidine-analbite relations are stable, where-

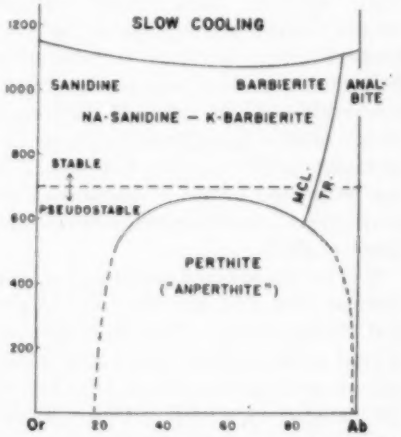

FIG. 9.—Phase relations of alkali feldspars that have been cooled slowly enough to unmix with respect to the alkali ions but not slowly enough to order the Al and Si ions.

as the diagram below the temperature of diffusive transformation (microcline-sanidine and albite-analbite) is pseudo-stable.

PHASE RELATIONS IN THE SERIES MICROCLINE-ALBITE

It seems desirable to discuss in a qualitative way the phase relations in the microcline-albite series, the ordered forms of the alkali feldspars. The temperatures used here are approximations and subject to future correction. The facts and assumptions necessary for this discussion are listed below.

1. Bowen (1913) determined the melting point of albite as 1,100° ± 10° C. Greig and Barth (1938) reported it as 1,118° ± 3° C. This later value was confirmed by Schairer and Bowen (1947). Greig and Barth discussed the difference between Bowen's value and theirs and suspected that the material used by Bowen was contaminated. Greig and Barth's determination was with synthetic material and therefore with the high-temperature modification (analbite), whereas Bowen used natural Amelia County albite, that is, the low-temperature modification. A measurable difference in melting points of the two is conceivable. Albite can and does exist pseudo-stably (superheated) above approximately 700° C., and, therefore, if it can be melted before it transforms, its melting point should be below that of the stable analbite.

2. The incongruent melting of potash feldspar was first described by Morey and Bowen (1922). There is no experimental evidence that shows any difference in melting temperatures for any of the varieties of potash feldspar. In the last paper on this matter (Schairer and Bowen, 1947, p. 200) it is stated: "The tiny crystals of potash feldspar obtained in melts in these studies did not permit a decision as to whether they are orthoclase or microcline." One can be quite sure that the decomposition temperature of 1,150° ± 20° C. reported by Schairer and Bowen is that of sanidine. The decomposition temperature of microcline (melting pseudo-stably) should be somewhat lower theoretically, analogously to the albite-analbite difference, and would not be expected to differ much from that of sanidine.

3. Investigation of cryptoperthites and moonstones (Laves, 1951*a*) indicates a continuous series of mix-crystals between microcline and albite. This was confirmed by an artificial preparation of microcline from a crystal of albite (Laves, 1951*c*), making use of the ease of alkali diffusion at high temperatures.

4. There is thus a strong indication that above a certain temperature there exists (at least pseudo-stably) a series of solid solutions between microcline and albite. At low temperatures, however, alkali feldspars show but very little solubility at either end of the series. Therefore, an exsolution curve similar to the one determined by Bowen and Tuttle (1950) for the sanidine-analbite series must exist also for the microcline-albite series.

5. The only way at hand to obtain the microcline-albite exsolution curve is to determine the lowest temperature of homogenization of *natural* microcline-albite perthites (or to determine the unmixing relations of such a homogeneous series). There seems little hope of using synthetic material, which is observed to belong always to the high-temperature series. Unfortunately, suitable natural material (microcline cryptoperthites) was not at hand for an experimental determination, but the following considerations can be used for a qualitative discussion.

Homogenization of moonstone from Grant County, New Mexico, takes place at approximately 50° C. below the temperature that would be indicated by the experimental exsolution curve of Bowen and Tuttle (1950). Their curve, it must be remembered, represents the exsolution relation of sanidine-analbite. There are not many explanations that can be put forward to explain the lower homogenization (or exsolution) temperature of the natural material. The effect of impurities can probably be ruled out. Aside from trace elements, anorthite ($CaAl_2$-

Si_2O_8) is the only material in solid solution (approximately 1–2 per cent anorthite). Potash feldspar does not form mix-crystals with anorthite; therefore, addition of lime to the system would tend to *raise* the unmixing curve, not lower it, as observed. The only other explanation obvious to the writer has to do with the influence of the Al-Si distribution in the $(AlSi_3)O_8$ framework. Natural feldspars should be expected to have a higher degree of order than the synthetic analogues. It is therefore considered probable that there are temperatures at which miscibility in the ordered microcline-albite series is greater than in the disordered sanidine-analbite series. This is merely another way of saying that the upper temperature of unmixing is lower for the ordered compounds.

6. Assuming this to be true, there are four configurations of the two exsolution curves that can exist, the two extremes being illustrated by figure 10. The lower exsolution curve exhibited by the low-temperature modification is not so unusual as it might appear at first sight. It is common for the low-temperature modification of silicates to take more in solid solution than does the high-temperature modification. See, for example, the system $NaAlSiO_4$-$KAlSiO_4$ (Bowen, 1917) and the subsolidus relations in the system $Mg_2Si_2O_6$-$CaMgSi_2O_6$ (Atlas, 1952). Of the two diagrams in figure 10, the writer considers *A* to be the most probable one.

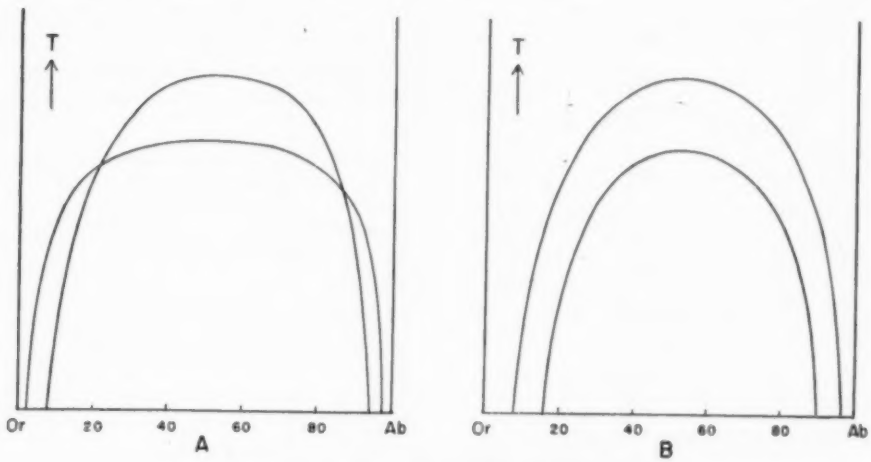

FIG. 10.—Hypothetical relations of exsolution curves

On the basis of the arguments in statements 1–6, the phase relations in the system microcline-albite are proposed in figure 11. The system is stable at low temperatures, but pseudo-stable (superheated) at high temperatures.

THE STABLE-PHASE DIAGRAM

To obtain the stable-phase relations, figures 9 and 11 must be combined. Before doing so, an important point must be discussed.

The work of Bowen and Tuttle (1950) showed the analbite-albite transformation temperature to be approximately

700° C. (The authors recognize the possibility that the transformation may take place at somewhat lower temperature, but they use the value of 700° in their phase diagram.)

Nothing, however, is known about the actual transformation temperature microcline ⇌ sanidine,[6] but most writers who have considered the alkali feldspars place it at approximately 700° C. on the basis of petrological observations (650° C. by Mäkinen, 1917; 650° C. by Alling, 1926, p. 596; "above 700° C." by Spencer, 1930, p. 363; about 750° C. by Spencer, 1938, p. 91; about 700° C. by Oftedahl, 1948, p. 68; approximately 675° C. by Barth, 1951). Realizing that the average opinion does not necessarily approach the truth, the writer nevertheless used the "average" value of 700° C. as a probable temperature. Reasoning on a structural basis would also indicate that this temperature is of the right order of magnitude. The chief difference between the molten and the crystalline states in the alkali feldspars is in the arrangement of the $(AlSi_3)O_8$ framework. The Na and K ions are both singly charged, and, in addition to their very small contribution to the total structural energy, the difference between them is quite small. It might, therefore, be expected that the difference in melting points of the two alkali feldspars would be rather small, as it is: 1,120° C. for $NaAlSi_3O_8$ and 1,150° C. (incongruent) for $KAlSi_3O_8$. For similar reasons one would expect that the microcline-sanidine and albite-analbite transformation temperatures would not differ much, assuming the transformation mechanism to be as here considered. The value of 700° C., which is the upper limit determined by Bowen and Tuttle (1950) for albite-analbite, is therefore considered reasonable for microcline-sanidine.

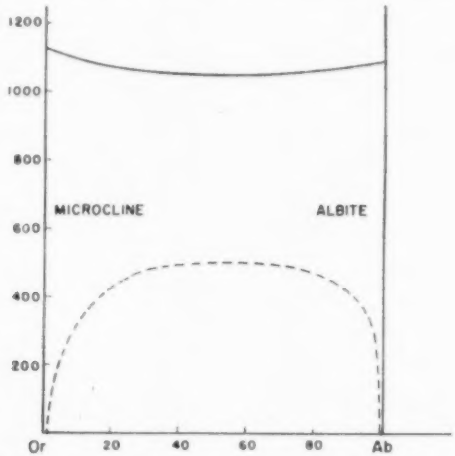

FIG. 11.—Probable relations in the microcline-albite series

[6] As the manuscript is being written, transformations from microcline to sanidine have been successfully carried out hydrothermally by my colleague, J. R. Goldsmith. Experimentation is still in progress, and indications are that the transformation takes place not far from 700° C. These experiments conclusively disprove the suggestion put forth by Eskola (1951) that microcline might be the high-temperature modification of potash feldspar.

Views expressed in the literature vary on transformation relations within the alkali feldspar series. The question of relations within the complete series has not been discussed, but there have been opinions expressed on the transformation near the potash side and near the soda side. Mäkinen (1917) thought that the transformation field rose with temperature from microcline-sanidine as the soda content increased (fig. 2, first part of paper, p. 437). On the soda side, Bowen and Tuttle (1950) showed the curves dropping sharply from the albite-analbite inversion point (fig. 5, first part of paper, p. 437). The writer believes that virtually

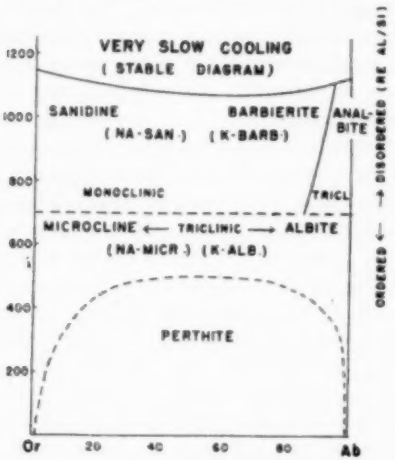

FIG. 12.—The probable phase diagram

complete solid solution exists between microcline and albite, as well as between sanidine and analbite, at temperatures near inversion. In addition, the transformation, here considered a diffusive one, would not *require* a two-phase field, as drawn in earlier diagrams. On the basis of the arguments used above with respect to inversion temperature of the end-members, a near-horizontal line is postulated to depict the inversion relations (fig. 12). It is, of course, possible that a two-phase field does exist and that it could be one of the two types shown in figure 13.[7]

An argument could be put forth against the stable-phase diagram (fig. 12) on the basis of the fact that microcline-albite mixed crystals in the center of the compositional range have not been ob-

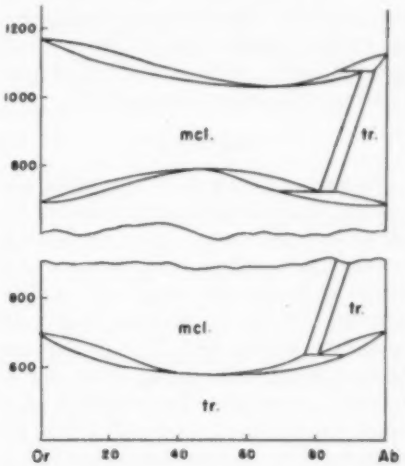

FIG. 13.—Two-dimensional field representatives of the transformation curves drawn in fig. 12.

served, whereas corresponding compositions are quite common in the high-temperature series. This objection can be met by considering that the time-consuming process of ordering Al and Si would require holding the material for a considerable time at elevated temperature and/or cooling very slowly. This process would then be expected to result in later unmixing because of the slowness

[7] Oftedahl (1948, p. 68) shows horizontal lines for the inversion at both ends of the diagram, reproduced in fig. 4 of the first part of this paper, but at different temperatures. Barth (1952, p. 255) shows a falling inversion on the albite side and a horizontal line on the microcline side. In neither case is any explanation given.

of the entire cooling process. A small rock mass that cooled rapidly would have insufficient time for the ordering process, and a large rock mass would cool slowly enough to produce later exsolution. Only under very exceptional circumstances could these microcline-albite solid solutions be observed. Such circumstances would be quenching of a mass held for very long periods below the transformation curve, or relatively short heating (as in contact metamorphism by a small igneous body, such as a dike) of preexisting albite-microcline perthite.

PERTHITES

As a consequence of the foregoing discussion, the cooling of any member of the alkali feldspar series, under equilibrium conditions, must result in atomic movements such that the stable states characteristic of the temperature finally attained are approached. This process can be concurrent with cooling, or it can conceivably come later. Two different processes are involved, both of which are necessary for the establishment of equilibrium. These are (1) the ordering of Al and Si and (2) the unmixing of K and Na. Under nonequilibrium conditions either one can take place independently of the other; if one takes precedence over the other under these conditions, total composition might be an important factor (i.e., a 50:50 composition might tend to unmix before ordering, whereas a 90:10 composition might tend to do the opposite). Both processes are undoubtedly continuous ("gradual") and, therefore, formation of intermediate states might be expected. The early stage of unmixing, for example, might be expected to produce "phases" structurally modified from those of the stable series. In the following discussion it should be understood that the ordinary mineral names used may apply to some of these "modified" phases.

For example, some of the cryptoperthites here investigated show "albite" to be "exsolved." This albite, however, has an a-axis considerably smaller than that characteristic of pure albite (compensated by a larger value of the b-axis). If one attempts to apply Vegard's law (as was done on pp. 550–553) to these a-axis values for determination of composition, one obtains impossible Na-feldspar contents of the order of 110 per cent. The reason for this is as follows: If a homogeneous feldspar unmixes, it begins in such a way that within the "single crystal Al,Si-O framework" fluctuations in composition appear that are expressed in single-crystal X-ray photographs. Two sets of diffraction spots appear, one set characteristic of a K-rich material, the other of a Na-rich material. These spots show characteristic varying degrees of diffuseness, but a variation that is a function of the size, shape, and orientation of the "unmixed" areas. As long as the Al,Si-O framework remains common to both areas, these areas cannot be expected to exhibit lattice constants that the individually crystallized phases would normally exhibit. It would be expected that the structures of the two areas would influence one another and that, rather than sharp boundary lines, broad boundary zones with continuously varying properties would be formed. In this way the strain produced by the "unmixing" (without the creation of new surfaces) can be kept at a minimum. This subject will be treated in detail in a later paper.

On the basis of this knowledge it might be desirable to divide the perthites into two major groups. (1) *Heterogeneous perthites* (*normal perthites*) are mixtures of two phases, one K-rich, the other Na-

rich. The composition of the two phases is determined by the appropriate exsolution curve. Theoretically, the phases are mechanically separable. (2) *Quasi-homogeneous perthites*[8] are not two distinct phases but represent, instead, an assemblage of areas having compositions approaching those values given by the appropriate exsolution curve. They are, in a sense, solid solutions with compositional fluctuations that considerably exceed those found in a normal mix-crystal. There are no distinct boundaries between the areas of differing composition. Many (virtually all?) cryptoperthites fall into this category.

Some general features of X-ray photographs of several cryptoperthites will next be discussed briefly and interpreted qualitatively.

1. Single-crystal photographs of cryptoperthites that are monoclinic commonly show, in addition to the diffraction spots from the monoclinic structure (potash-rich), reflections elongated in the direction of the b^*-axis or in a direction [10l] with l approximately 6–8. These additional reflections commonly show intensity fluctuations, the intensity fading out from the "center" of the "streak." In quite a few cases the fluctuations in the streaks are so strong that, instead of being "elongated reflections," they appear as sets of sharp spots (5–12 spots in one streak have been observed) with spacings that indicate a periodicity within the crystal of the order of 100 Å, but somewhat variable from crystal to crystal (discussed briefly by the writer, 1951*a*). If one considers only the centers of those streaks or reflection sets, they can be related to triclinic material with lattice constants close to albite (in the case of elongation in the b^*-direction) or close to analbite (in the case of elongation in the [10l] direction). Similar effects were observed by Chao and Taylor (1940), Ito (1950), and Saalfeld (1952). The writer, however, does not agree with all the interpretations of these authors. Ito assumes the existence of several "albites" with differing lattice constants in a single perthite as the source of the reflection sets. This explanation is not in harmony with that of the writer.

In the lower part of figure 14 six monoclinic cells are shown with the normal b-axis length. If in such a monoclinic host areas of triclinic Na-feldspar are "exsolved," a strain is produced by the triclinic cells. This is shown in the upper row of triclinic cells in figure 14 (the triclinicity is exaggerated). The strain will obviously be lessened by repeated twinning, resulting in a b-axis periodicity greater than that of the original monoclinic cell. In figure 14 the periodicity of the triclinic material = b′(twin) = 6b. A periodicity of this sort would produce the sequence of reflections observed in the single-crystal X-ray photographs described above. It is to be expected that, in the equation $b' = nb$, the value of n would vary in different crystals as a result of variability of strain relations during "exsolution." It is also to be expected that n represents an average value and is not normally an integer. The fact that these extra reflections vary from streaks to sharp spots is easily explained by this model in the same way as the variation in distances between the spots from crystal to crystal.

It was observed also that all specimens with "exsolved" albite showed albite twin relations, whereas those with "exsolved" analbite showed pericline twin relations. The different lattice constants of albite and analbite may well be responsible for the differing twin relations. The writer's interpretation of Chao and

[8] Term used by Oftedahl (1948, p. 67).

Taylor's (1940) data is that their samples followed the same rule.

2. The strong influence of the host on the "exsolved" material becomes apparent if one analyzes the mutual orientation of the two. In figure 14 the two materials were related by coincidence of the (010) plane (albite twin law). For simplicity, the *a*- and *c*-axes were also drawn in coincidence. In reality, this latter coincidence cannot exist, as β differs in the two materials. Böggild (1924) and Chao and Taylor (1940) have shown that these two axes do not coincide in perthites and cryptoperthites.

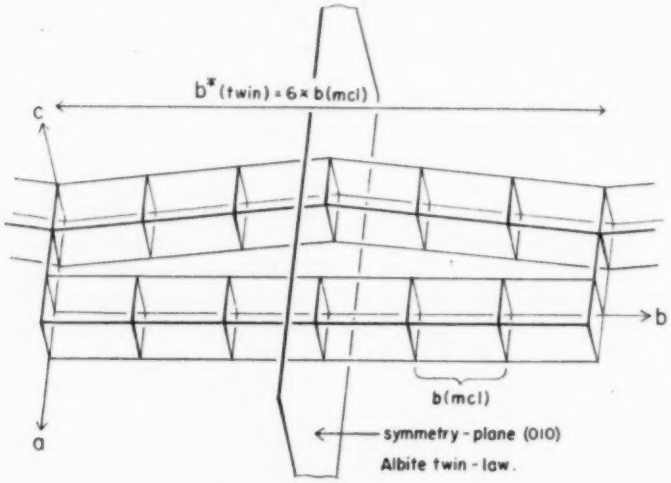

FIG. 14.—Twinning superstructure in cryptoperthites (schematic)

An infinite number of possible mutual orientations exist. Two orientations are shown in figure 15. In this figure the difference in β is neglected ($\sim 116°0'$ for K-feldspar, $\sim 116°40'$ for Na-feldspar). The difference in *a*-axis lengths of the two materials is quite significant, however. If *a* and *c* coincide, as in figure 15, *A*, the [101] directions cannot coincide. In like fashion, if the [101] directions coincide, the *a*- and *c*-axes cannot (and so on for other directions). There is one mutual orientation, however, that is unique, in that the sum of the values of all deviations is at a minimum and approaches zero. This is shown diagrammatically in two dimensions in figure 15, *B*. As far as the writer's experience goes, the cryptoperthites seem to follow this scheme of minimal deviation of the lattice rows. An example of a quantitative determination of these crystallographic orientation deviations is shown in figure 16. It should be noted that in this arrangement two directions are observed to coincide for both materials and, of necessity, form an angle of 90°. In the cryptoperthite shown in figure 16 these directions are very close to [301] and [106]. It is of interest to note that in moonstones the plane of schiller as reported in the literature varies from $(\bar{6}01)$ to $(\bar{8}01)$. The [106] direction is parallel to the $(\bar{6}01)$ plane, and [301] is normal to it, indicating a relation between these structurally controlled directions and the schiller "planes."

3. It is also of interest to observe the

behavior of the "exsolved" areas when heat-treated. Cryptoperthites with exsolved *analbite* can be homogenized quite rapidly at temperatures above the solvus (in hours at 700° C., and within minutes at 1,000° C.). Below the solvus, heat treatment rapidly produces a change of composition indicated by a change in the lattice constants. The compositions approached are those indicated by the solvus experimentally determined on synthetic feldspars by Bowen and Tuttle (1950), taking into account the restrictions discussed on page 562. It was observed also that heat treatment at 510° C. changes the lattice angles α and γ of the "exsolved" analbite to virtually 90°, producing virtual monoclinic symmetry. This was determined on moonstone from Grant County, New Mexico ($\sim Or_{58}Ab_{42}$), and on cryptoperthite from East Africa[9] (material 5 of table 2). This change would be expected on the basis of phase relations presented in figures 7–9, especially if the displacive transformation, triclinic-monoclinic, is reinforced at elevated temperatures by the monoclinic environment of the K-rich host.

The reverse situation, that is, influence of a triclinic host on a guest that should be monoclinic, could also be observed on other crystals (optically triclinic) from the same hand specimen of rock from which the optically monoclinic cryptoperthites discussed above were obtained. Although most of the phenocrysts of this rock were "monoclinic," some could be recognized to be "triclinic," even macroscopically, by the presence of albite twin lamellae. This material was discussed in the section on phase relations in the sanidine-analbite series in another connection, and the lattice constants of the homogenized material are listed under "4" in table 2. Before heat treatment the material was a triclinic cryptoperthite with α^* and γ^* that differed but little (less than 15′) from the α^* and γ^* of the homogenized material. The photographs taken before the heat treatment showed three different sets of spots: (1) those of a K-rich material, (2) those of a Na-rich material, and (3) those of a material with

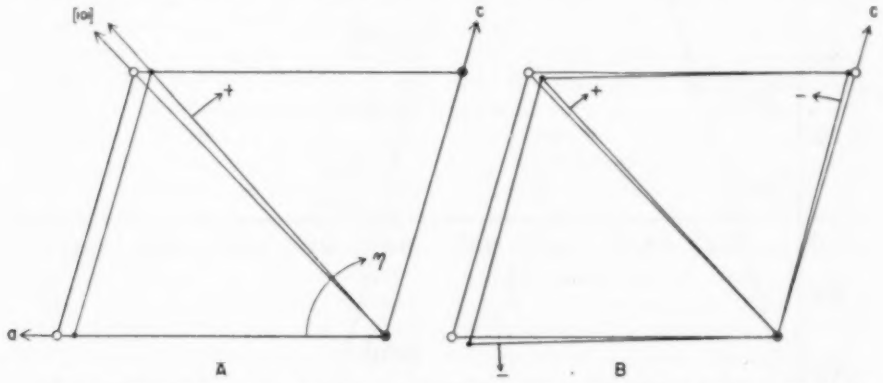

FIG. 15.—Schemes of mutual orientation of the exsolved materials in cryptoperthites. The arrows indicate the directions (+, −) of deviations of the lattice rows between the two materials. The solid circles are for albite, the open circles for the orthoclase lattice.

[9] Kindly supplied by N. L. Bowen for further check of the originally puzzling observation on the Grant County moonstone.

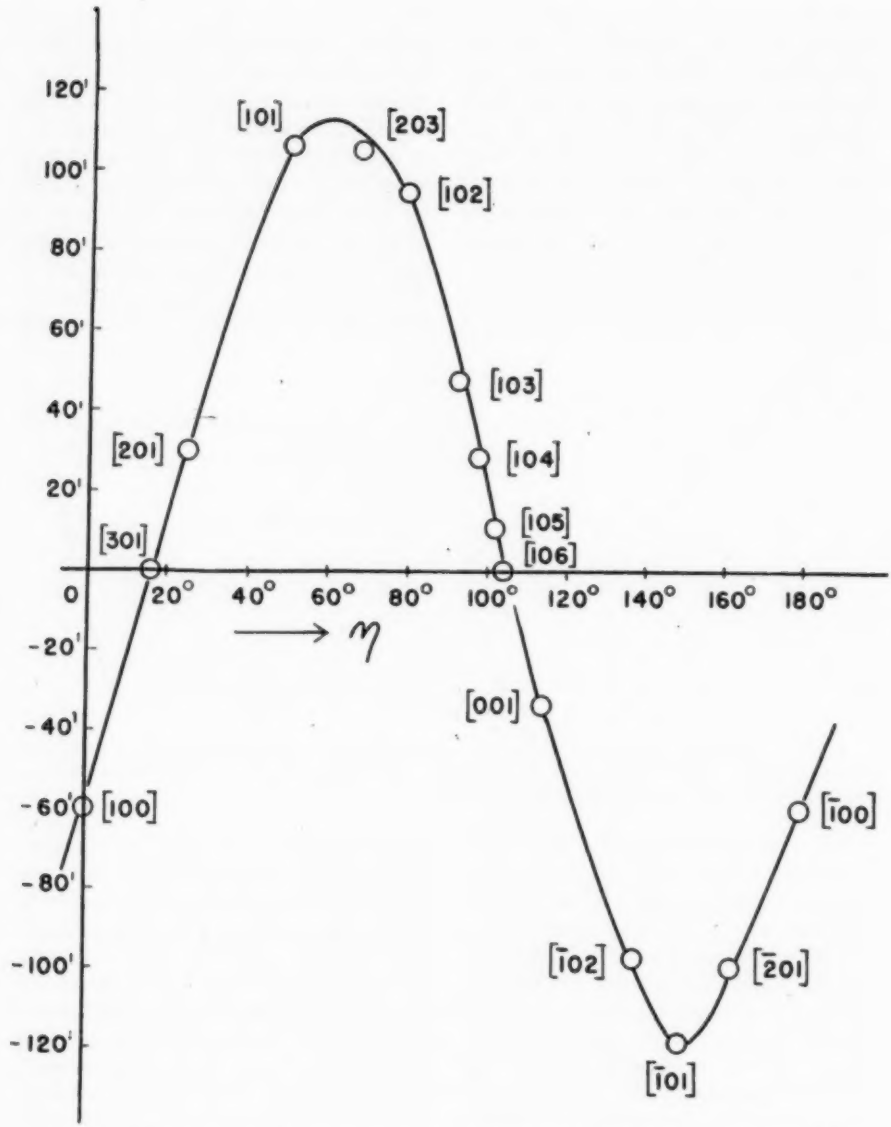

FIG. 16.—The experimentally (Buerger precession method) determined deviations of several lattice rows of albite relative to orthoclase in a cryptoperthite, plotted as a function of the azimuth η (the direction of the a-axis defined as having the azimuth $\eta = 0°$). A cryptoperthite kindly supplied by C. Oftedahl was used and described in his paper (1948) on p. 50 as sample P1.

intermediate composition. This third set of spots is virtually the same as the single set produced after homogenization of this crystal. All three sets have the same a^* and γ^*, which shows the strong influence of the triclinic host on the geometry of the "unmixed" areas in such a "quasi-homogeneous" material. It might be worth noting that the triclinic material of sample 4 was not completely unmixed, whereas the monoclinic material of sample 5 was.

These observations allow one to draw some conclusions on the temperature at which exsolution took place in this particular rock. On pages 555–556 it was shown that the homogenized sample 5 showed a monoclinic $\rightleftharpoons$ triclinic displacive transformation at 55° C., whereas sample 4 transforms at 115° C.[10] It appears, therefore, that the development of the cryptoperthitic state took place in sample 4 above 55° C. and in sample 5 below 115° C. It seems safe to conclude that the "exsolution" process did not take place before the rock had cooled to approximately 100° C.

4. Cryptoperthites with exsolved *albite* behave in a strikingly different way when heat-treated. They cannot be homogenized in a short time. The materials investigated remained unmixed after heating to 1,060° C. for 2 weeks. The writer's explanation is that in these cryptoperthites the Al-Si distribution of the exsolved albite areas is so different from that of the host that the short heatings that are sufficient to interchange the alkali ions are not sufficient to produce the same degree of Al-Si disorder in guest and host. The albite areas are assumed to be highly ordered, whereas the K-rich areas are assumed to be in a state corresponding to "common orthoclase" in which a considerable degree of disorder is still prevalent. Therefore, the only effect of heat treatment that is observable in a short time (hours at 700° C., minutes at 1,000° C.) is the development of a distribution equilibrium of Na and K ions between the two Si-Al framework areas which differ in the degree of disorder (i.e., Na and K equilibrium is established between albite and orthoclase). This distribution equilibrium is a reversible function of temperature. At temperatures near those of melting the unmixed areas approach one another in composition. This difference becomes smaller after long heating periods until the areas eventually become indistinguishable, homogenization taking place only when the same degree of disorder has been induced in both materials. This explanation accounts for the observation of Kôzu and Endô, who in 1921 demonstrated the existence of two kinds of cryptoperthites, one kind being described as having a "memory." In the writer's opinion the "memory" is a consequence of the reluctance or, better, extreme sluggishness involved in disordering the ordered Al-Si framework of the albite areas. If analbite is exsolved, no "memory" is involved, because the degree of disorder in these areas is akin to that of the K-rich areas and random alkali-ion distribution takes place readily at elevated temperatures.

[10] The "transformation temperature" of sample 4 before homogenization (at 800°) was approximately 165° C. An analogous curve, as shown in fig. 6, could be obtained. This again illustrates the influence of the different "quasi-phases" on one another in a "quasi-homogeneous" mix-crystal.

ON THE MEANING OF THE NAME "ANORTHOCLASE"

No other mineral of importance appears to play such an ambiguous role in mineralogical and petrographic literature as "anorthoclase." Numerous attempts have been made to define and characterize this mineral, and virtually every pub-

lication and textbook considers it in a different way. This is not surprising when one realizes that a specific mineral "anorthoclase" does not exist if the use of the word "mineral" is restricted to its generally accepted definition.[11]

Most recent textbooks describe anorthoclase as a sodium-rich variety of microcline (for example, see Kraus, Hunt, and Ramsdell, 1951). Winchell, with this point of view, proposed a microcline-anorthoclase-analbite series. In his original paper on the subject (1925) he used the anorthoclases of Förstner (1884) to complete the optical data for his series. He neglected, however, many of Förstner's other observations on the same material. For example, Förstner determined, among other things, values of α on analyzed specimens, three of the values being 91°1′, 91°42′, and 92°3′. He described the triclinic-monoclinic inversion and thus showed clearly the variability of the crystallographic data. In Winchell's book (1951), however, anorthoclase is described on page 311 as having $\alpha = 91°19'$, giving the impression of a well-defined value for a fixed mineral. With respect to inversion, he stated: "Anorthoclase is stable to its melting point." On page 298 he called it "metastable at ordinary temperature," and on page 265 he stated α to be 90°30′. Inconsistencies or uncertainties such as those in this one book are common in the entire literature, as far back as the date of coinage of the term "anorthoclase."

Rosenbusch (1885) first proposed the name (*Anorthoklas*) and defined it as follows:

> Die Reihe der triklinen Kali-Natronfelspäte, zu deren wichtigsten Eigenschaften gerechnet werden muss, dass sie einen scheinbaren Spaltwinkel P:M besitzen, der kaum oder nicht nachweisbar von einem rechten abweicht und doch davon abweichen muss, soll als die Reihe der *Anorthoklase* bezeichnet werden im Gegensatz zu den nachweisbar schief spaltenden Plagioklasen. ["The most important property of those soda-potash feldspars that will be called the 'anorthoclase series' is the cleavage angle P:M that 'appears' not to deviate from a right angle, yet which must do so. This series differs from the plagioclase which has obviously oblique cleavage."]

The fact that the cleavage angle P:M appears close to 90° was believed by Rosenbusch to be due, at least in part, to very fine twinning, such that (001) and (010) "appear" as a planar cleavage surface.

Rosenbusch based his definition on several descriptions of such feldspars, mainly those of Förstner (1883) and Brögger (1882). On the basis of the known compositions he placed this series in the range $Or_{33}Ab_{67}$–$Or_{18}Ab_{82}$, with a variable additional anorthite content, An_4–An_{33} per $100(Or + Ab)$. He also listed as properties the variability of the geometrical angles (α, β, γ) and the monoclinic-triclinic inversion that takes place on heating, but he assumed that the transition takes place only in the Ca-poor material. Anorthoclase was treated in his classification in a supplement to the chapter on the plagioclases, microcline being treated separately. In the fourth edition of his book (1904, p. 326) he stated that perthitic structure (virtually always seen in microcline) had never been observed in anorthoclase.

Brögger (1890, pp. 539, 563; 1897, pp. 12–14) argued strongly against the name "anorthoclase" and described (1882, pp. 262, 292) as "soda-microcline" (*Nathronmikroklin*) the very same specimens that Rosenbusch two years later called "an-

[11] "A mineral may be defined as a substance occurring in nature with a characteristic chemical composition and usually possessing a definite crystalline structure, which is sometimes expressed in external geometrical forms or outlines" (Kraus, Hunt, and Ramsdell, 1951, p. 6).

orthoclase." He felt this name to be superior because it better expressed the "close" relation of this material to microcline. That this relation is erroneous was discussed in the sections on phase relations in the sanidine-analbite and microcline-albite series. Therefore, the term "soda-microcline" as a synonym for "anorthoclase" should be dropped.[12]

A third point of view, set forth by Oftedahl (1948, pp. 56–58), is that inhomogeneity of a cryptoperthitic nature is an important characteristic of anorthoclase. He did not consider the possibility of a homogeneous triclinic Na-rich alkali feldspar, but ascribed the triclinic nature to unbalanced exsolution of triclinic material in a monoclinic (K-rich) host. It is the writer's experience that this is occasionally, but not always, true. In view of the fact that most petrologists think of sanidine as having a small axial angle, Oftedahl stated (p. 56):

> Thus the term sanidine may overlap what is mostly called anorthoclase (soda-rich alkali feldspar, more or less deviating from monoclinic symmetry). By studying the previous literature, one gets the impression that the symmetry relations are too much stressed in relation to other optical relations. The difference in symmetry between microcline and orthoclase is very marked, but among the alkali feldspars I have found nearly monoclinic members, and feldspar with a distinctly triclinic symmetry. With such a gradual transition the symmetry relations are unusable as characteristics.

Oftedahl defined three series (p. 67):

> 1. The sanidine series, monoclinic, homogeneous and possibly continuous. 2. Orthoclase-cryptoperthite series. Quasi-homogeneous, from sanidines by exsolution. Discontinuity from 70 to 100 Ab. Symmetry mostly monoclinic, but grades into triclinic symmetry (anorthoclase). 3. Microperthite series, discontinuous. . . .

[12] Oftedahl (1948, p. 57) also advocates dropping this term.

Alling (1926, p. 600) neglected the symmetry relations even more than did Oftedahl and based his definition of anorthoclase exclusively on the chemical composition. He stated:

> A common conception of the term "anorthoclase" is that it is a potash-bearing soda feldspar, presumably a single-phase system. Winchell still accepts this view. I advocated a modification in meaning, in spite of precedent of the original use of the term. "Anorthoclase" in my sense is α, β, or γ[13] feldspar composed at least of the two minals ranging from $Or_{70}Ab_{30}$ to $Or_{30}Ab_{80}$, which exist as a single phase system. If the views here suggested are correct, then γ anorthoclase is monoclinic, β anorthoclase may be monoclinic and triclinic, while α anorthoclase is triclinic. Only γ anorthoclase is stable, and that above 900°–950° C. range in temperature. I believe that there is sufficient evidence to entertain the view that many perthites are (1) the result of decrease in solubility attendant upon inversion, and (2) due to any form of anorthoclase, a single-phase system, becoming a two-phase system by exsolution.

This point of view was retained in principle in Alling's book (1936).

Summarizing the views of previous writers, the following conceptions of "anorthoclase" appear to be currently held:

1. *Brögger:* The similarity of anorthoclase (including its triclinic nature) to microcline is stressed.

2. *Rosenbusch:* The similarity of anorthoclase (including its triclinic nature) to plagioclase is stressed.

3. *Oftedahl:* The nonexistence of a homogeneous triclinic phase is stressed, anorthoclase being a cryptoperthite that may or may not be triclinic. The apparent symmetry is governed chiefly by the balance or lack of balance in the fine twinning of the unmixed material.

[13] Not crystallographic symbols. Alling proposes three (occasionally four) feldspar series: (α) microcline-analbite, (β) orthoclase-albite, (γ) sanidine-barbierite.

4. *Alling:* Anorthoclase may or may not be homogeneous; it may or may not be or appear to be triclinic. An essential composition range of $Or_{70}Ab_{30}$ to $Or_{20}Ab_{80}$ is defined.

The foregoing discussion gives an indication as to why the writer has not used the term "anorthoclase" in the proposed phase diagrams.

PROPOSED NOMENCLATURE

The following nomenclature is based principally on an attempt to give an unequivocal designation to every homogeneous alkali feldspar, whether that feldspar be stable, pseudo-stable, or unstable. Thus it was considered necessary to distinguish with different names those states of the pure end-member compounds that can be distinguished by structural type. Whether or not such compounds actually exist in the chemically pure form was not deemed essential to the classification. In a somewhat modified way, these states correspond in principle to Alling's "minals" (1926, p. 592). Traditional terminology is followed as far as possible.

1. The monoclinic *sanidine-barbierite* series is stable at high temperatures. However, a pure monoclinic $NaAlSi_3O_8$, disordered with respect to Al-Si, has not yet been demonstrated. The series may be divided on a compositional basis into *sanidine*, *Na-sanidine*, *K-barbierite*, and *barbierite*.

2. Pure $NaAlSi_3O_8$, on the soda side of the sanidine-barbierite series, is a triclinic modification that is called *analbite* (Winchell's name, more clearly defined). At low temperature, analbite is capable of taking quite a bit of potash in solid solution (fig. 8), and it is the analbite series that exhibits the displacive triclinic $\rightleftarrows$ monoclinic transformation. The K-rich members can be called *K-analbite*, and are the triclinic low-temperature (although pseudo-stable at low temperatures) equivalents of the monoclinic barbierites.

3. *Microcline* and *albite* are the alkali feldspars stable at low temperatures. The series *microcline-albite* was discussed previously (fig. 11). It can also be divided compositionally into *microcline*, *Na-microcline*,[14] *K-albite*, and *albite*.

4. On the potash side of the alkali feldspar series additional designations are necessary to characterize the materials that show special optical, X-ray, and morphological features. The names *orthoclase* and *adularia* have long been in use, and the writer's opinions on the unstable nature of these minerals was discussed in the first part of this paper (pp. 445–449). Because usage of the word "orthoclase" is not always uniform, the writer suggests the following definition: *Orthoclase* is potash feldspar that is or appears to be monoclinic and includes *sanidine*, "*common orthoclase*," and *adularia*. "Common orthoclase" was discussed in the first part of the paper (p. 446), and adularia includes the *monoclinic* and (slightly) *triclinic* material.

5. Members of the sanidine-barbierite series would exsolve under equilibrium conditions at low temperatures. If, however, this material were held for some time at a temperature that was above the solvus but below the monoclinic-triclinic diffusive transformation temperature (*ca.* 700° C.?), and then cooled rapidly enough so that exsolution did not take place, this material might be expected to be observed as a member of the microcline-albite series (an ordered or partially ordered analogue of the sanidine-barbierite series). If the time was insufficient to transform this material to the ordered

[14] Not to be confused with the older term that has been used as a synonym for "anorthoclase."

state, it might persist in an "unstable" intermediate state or states that might be considered characteristic of "common orthoclase." Such material may be called *Na-common orthoclase*. Analogous intermediate states between analbite have not been observed in natural crystals.

6. The terms *perthite*, *antiperthite*, *macro-*, *micro-*, and *cryptoperthite*[15] are all here used with their "traditional" meanings. Tuttle (1952, p. 114) further splits cryptoperthites into *X-ray perthites* and *sub-X-ray perthites*, the latter being a perthite in which the (201) powder diffraction line of the exsolved material cannot be observed with an X-ray spectrometer.[16] Perthites (cryptoperthite?) which exhibit schiller are called *moonstone*. A term *Anperthite* could be used for a material with exsolved analbite.

7. In many respects the state of the material that forms the perthite is as significant as the size. This has been discussed in detail in the section on perthites. If one knows (or suspects) the state of the material, it can be described by terms such as *orthoclase-analbite cryptoperthite* (orthoclase, in this case either sanidine or "common orthoclase" as host, with exsolved analbite as guest); *albite-microcline antiperthite* (microcline in an albite host); and *analbite-tr. $KAlSi_3O_8$-anticryptoperthite* (actually observed by the writer [see p. 567]; analbite as host with triclinic K-feldspar[17] exsolved as guest). These are but a few of the possible combinations.

8. If it should be necessary to subdivide the sanidines (see p. 449 and p. 554), it is suggested that the terms "high-sanidine" and "low-sanidine" be avoided. The names "sanidine (high)" and "sanidine (low)" are suggested.

9. If one determines a feldspar to be in the composition range defined by Alling ($Or_{70}Ab_{30}$-$Or_{20}Ab_{80}$) but does not know enough about it or does not care to place it in one of the above-listed eight general categories, it is perhaps best called *anorthoclase*.

It should be pointed out that the proposed nomenclature has the advantage of ready extension into the ternary system of feldspars; in portions of the ternary system a split between alkali and plagioclase feldspars is artificial.

TERNARY FELDSPARS

Alling (1921) assembled compositional data on all feldspars and plotted them in a concentration triangle (Or, Ab, An). From these data one can obtain a fairly good idea of mutual solubility. His work indicates that at high temperatures, in addition to the Ab-An series, homogeneous feldspars of compositions lying to the right of curve *Or-D-C'-An* (fig. 17) exist. Franco and Schairer (1951) have shown that only one feldspar liquidus surface is obtained in dry melts in the system Or-Ab-An. Quenching undoubtedly prevents exsolution or ordering from taking place. However, the displacive transformation of the alkali feldspars shown in figures 7 and 8 is not prevented by quenching, and the extension of these compositions into the ternary system would be expected to produce ternary feldspars that also show this transformation. The fact that crystals 2 and 3 (table 2), used for the determination of the displacive transformation, contained ~2 and 6 per cent An, respectively, shows this to be the case. The Ca-content neces-

[15] The writer feels that the limitation of cryptoperthite to exsolved material in the size range 1–5 μ (see Tuttle, 1952, p. 114) is one that would defy the determination by many investigators. The term is perhaps best used for any submicroscopic perthite.

[16] The writer feels that this does not necessarily mean that this or other reflections might not be observed by single-crystal techniques.

[17] Not microcline, but K-feldspar "forced" to be triclinic by influence of its triclinic host.

sary to inhibit the transformation is unknown as yet. The description of the Ca-rich alkali feldspars and of the K-rich plagioclases that are typical of the rhombporphyries (see Oftedahl, 1948, and Osten, 1951, and literature there quoted) and related rocks supports the following conclusions:

1. The field of monoclinic feldspars (sanidine-barbierites) extends into the ternary system.

2. Point *A* (fig. 8), which separates the monoclinic barbierite from the triclinic

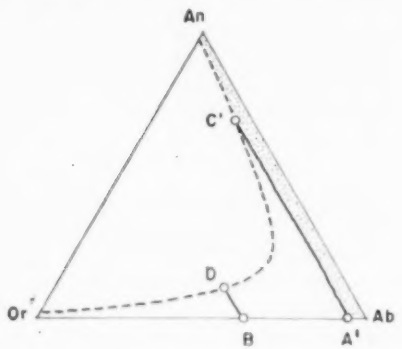

FIG. 17.—Base of the ternary diagram of rapidly cooled feldspars as presented in fig. 18.

analbite, becomes a line *A–C* (fig. 18) separating monoclinic *Ca-barbierites* from triclinic *anplagioclases* (term derived from analbite). This line, the position of which is as yet unknown, is schematically drawn in figure 18.

3. Point *B* (fig. 8), which separates the monoclinic K-barbierites from the triclinic K-analbites at room temperature, becomes a line *B–D* (figs. 17, 18) separating the *calcic K-barbierites* from the *potassic anplagioclases*.

4. The curve *A–B* (fig. 8) becomes a surface *A–B–D–C* (fig. 18) between the monoclinic sanidine-barbierite series and the triclinic anplagioclase series.

Keeping in mind that these feldspars are stable as homogeneous crystals only at high temperatures (and should exsolve at low temperatures, as do the sanidine-barbierite series), we may eventually be able to approach an understanding of the puzzling features considered by many authors (see Oftedahl, 1948, and references therein). The "appearance" (symmetry) of the ternary feldspars may be governed largely by the respective values of the transformation temperature and the temperature at which exsolution takes place. Exsolution can take place while the crystal is still monoclinic (before the displacive transformation), and the feldspars therefore appear monoclinic, or it can take place after the transformation, in which case the feldspars appear triclinic. Thus a careful investigation of symmetry, exsolution, and temperature relations may be useful as a geologic thermometer.

X-ray investigation of such ternary feldspars in the original state and after heat treatment has revealed that the "exsolution surface" which corresponds to the curve or "solvus" of the binary alkali feldspar system seems to be rather complex. In many cases three different exsolved feldspars are observable, the compositions of which are close to albite, orthoclase, and oligoclase. An "exsolution channel" connecting the alkali feldspar perthites with the plagioclase peristerites is therefore suggested. That the peristerites (approximately $Ab_{90}An_{10}$-$Ab_{85}An_{15}$) are composed of two feldspars with compositions close to albite and oligoclase has been reported by the writer (1951*a*). A detailed presentation of the X-ray work on these feldspars will be published later.

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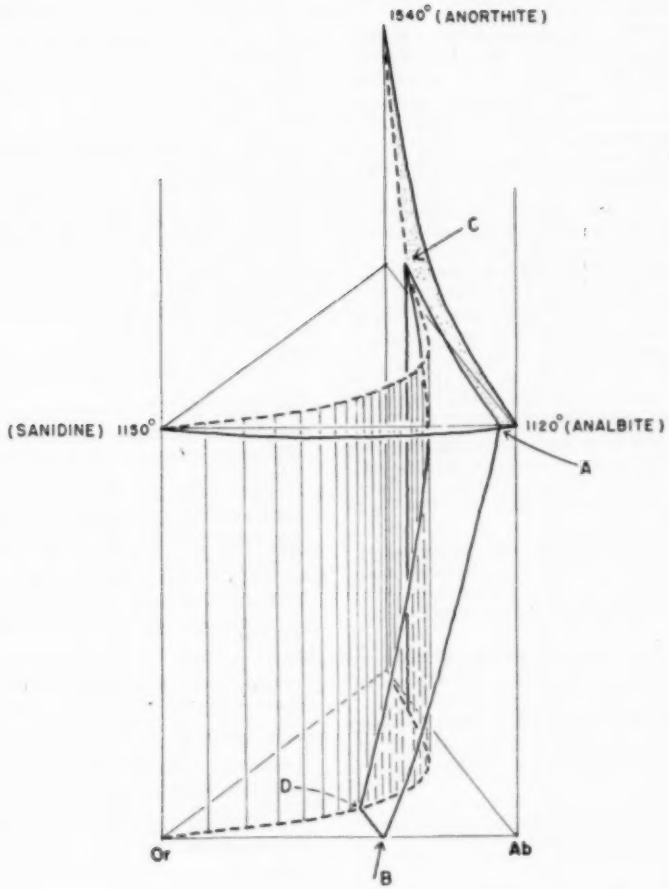

FIG. 18.—"Three-dimensional" diagram, showing a possible extension of the alkali feldspar diagram as presented in fig. 8 into the ternary system K-Na-Ca-feldspar. The front side of this diagram is identical with fig. 8.

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[18] These references are in addition to those listed in the first part of this paper (pp. 449–450).

PLEISTOCENE EVOLUTION OF THE OHIO AND WABASH VALLEYS[1]

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ABSTRACT

Studies of drift thickness and character currently in progress in Indiana indicate changes in drainage that have occurred in Indiana and some adjoining states since the beginning of the Pleistocene epoch. After late Tertiary (Lexington) peneplanation the master-drainage line, the Mahomet-Teays Valley system, became entrenched about 200 feet below its former level before it was ponded by Nebraskan ice and diverted into a new course. During the Parker cycle, which was interrupted by glaciation, the present physiographic features of southern Indiana and their now buried extensions to the north became recognizable. Static rejuvenation then was a major factor in erosion of the "Deep stage" of the Ohio Valley. Reversal of the pre-Illinoian Miami Valley probably resulted from this glaciation also. Further ponding of the Mahomet by Kansan ice and diversion of its drainage down the Wabash Valley is suggested as the origin of the "Deep stage" along that stream. Present drainage in central Indiana is a result of modifications produced by Tazewell and Cary glaciers.

INTRODUCTION

Differences of opinion concerning preglacial and interglacial stream courses in the states drained by the Ohio River and its tributaries from the north have long existed. Recently, a report on the bedrock topography of Illinois has been published, in which the evolution of present drainage is discussed and figured (Horberg, 1950, fig. 21). A similar study for Indiana currently in progress by the writer shows the existence of many previously unrecorded buried valleys and supplies additional data for interpretation of the erosional history of the region.

This report presents the writer's views of the sequence of changes in the drainage basins of the Ohio and Wabash valleys that resulted from Pleistocene glaciation. Approximately 4,000 water- and oil-well records in central Indiana have been collected and studied, the thickness of glacial drift plotted on county maps, and isopachous contours drawn from these data and outcrop information. Refraction seismograph data and well logs for the northern fourth of Indiana have not been studied.

This investigation is part of a survey of the mineral resources of Indiana and is being made under the direction of W. D. Thornbury, chief of the Glacial Geology section of the Indiana Department of Conservation, Geological Survey.

PREGLACIAL DRAINAGE

MAHOMET-TEAYS VALLEY

The main preglacial river of Indiana was the Mahomet-Teays, which probably headed in the Piedmont of North Carolina (Stout and Schaaf, 1931, p. 671) and flowed across West Virginia, Ohio, Indiana, and Illinois (fig. 1). An early ice sheet caused the first derangement of the river by severing the part called the Teays Valley (Tight, 1903, pp. 50–57, pl. 9) from that downstream.

Thornbury (1948, p. 1359) suggested using the name "Kanawha Valley" for this preglacial drainage line. In the opinion of the writer the name "Mahomet," proposed by Horberg (1945, p. 349) for the lower part of the valley which crosses Illinois to its discharge point on the preglacial Mississippi, could appropriately

[1] Published with the permission of the state geologist, Indiana Department of Conservation, Geological Survey. Manuscript received October 19, 1951.

be used for the valley across Indiana as well (fig. 1). The entire preglacial river valley might well be designated as the "Mahomet-Teays system," but the shortened interglacial valley of Illinois, Indiana, and western Ohio (fig. 2) should be called the "Mahomet" to distinguish it from its longer pre-Pleistocene ancestor.

One of the major tributaries of the Mahomet-Teays system in Indiana appears to have headed in southeastern Indiana and flowed northwestward to join the Mahomet Valley near Lafayette (fig. 1). The name "Anderson Valley" is suggested for the course of this preglacial river, as Anderson, Indiana, is situated on the till plain above it.

OHIO VALLEY

The pre-Pleistocene Ohio River was a relatively insignificant stream, which probably headed on the Silurian limestones[2] that underlie the area of the Muscatatuck regional slope (fig. 3) southwest of Madison, Indiana. Leverett (1902, pp. 116–118), Fenneman (1916, pp. 118–119), Ver Steeg (1938, pp. 657–659), and others thought that the Ohio headed upstream from Cincinnati. They based their conclusions mainly on the fact that the Miami Valley bedrock floor has a southwest gradient, whereas it should have a northeast gradient if it flowed toward the Mahomet-Teays. Malott (1922, pp. 136–138) and Fowke (1925, p. 87) favored the view that the Ohio headed near Madison, Indiana, and that the Kentucky River flowed northeastward through the old Miami Valley, which emptied into the Mahomet-Teays.

Considerable evidence has been offered to support the hypothesis of a former divide near Madison, Indiana. The Ohio River Valley is extremely narrow there, and the adjacent upland is undissected (Fowke, 1925, p. 87). The upland on the Muscatatuck regional slope near Madison shows a lack of beveling toward the Ohio River, but beveling does occur along the Kentucky River and its supposed continuation northeast from Madison (Fowke, 1925, p. 89). Beveling toward the Ohio does occur below Madison (Malott, 1922, p. 138). Many of the streams which join the Ohio above Madison exhibit a barbed pattern, but those that enter the Ohio below the old divide show normal relationships (Malott, 1922, p. 138). "Lafayette" gravels are present in quantity along the Kentucky River and along the Ohio downstream from New Amsterdam at the mouth of Indian Creek in Harrison County (Leverett, 1902, pp. 111–112). The upland is so well preserved in the Madison area that it is remarkable that they are not now present there, if they ever were deposited (Malott, 1922, p. 138). The presumed southwesterly slope does not take into account the possibility of uplift to the north following glacial unloading (Malott, 1922, p. 138) or the possibility of interglacial reversal of drainage in southern Ohio.

In view of the fact that the Mahomet-Teays Valley system apparently skirted the Cincinnati uplift on the north instead of flowing directly across it, as does the present Ohio River, it would seem likely that other drainage also would have become adjusted to that structure. The parts of Kentucky, Ohio, and Indiana underlain by Ordovician limestones and shales constitute essentially a shallow topographic basin. The lower part of the Kentucky River lies entirely within this basin and enters the Ohio River above the probable old divide at

[2] J. B. Patton (personal communication) indicates that the formation which generally forms the divide is the Laurel limestone.

FIG. 1.—Principal preglacial drainage systems of Indiana and adjoining states. Compiled from studies by Leverett (1915, 1929), Malott (1922), Stout, Ver Steeg, and Lamb (1943), Fidlar (1948), McGrain (1949, 1951), Horberg (1950), and others.

FIG. 2.—Drainage modifications that resulted from Nebraskan glaciation. Ohio data after Stout, Ver Steeg, and Lamb (1943); Illinois data after Horberg (1950).

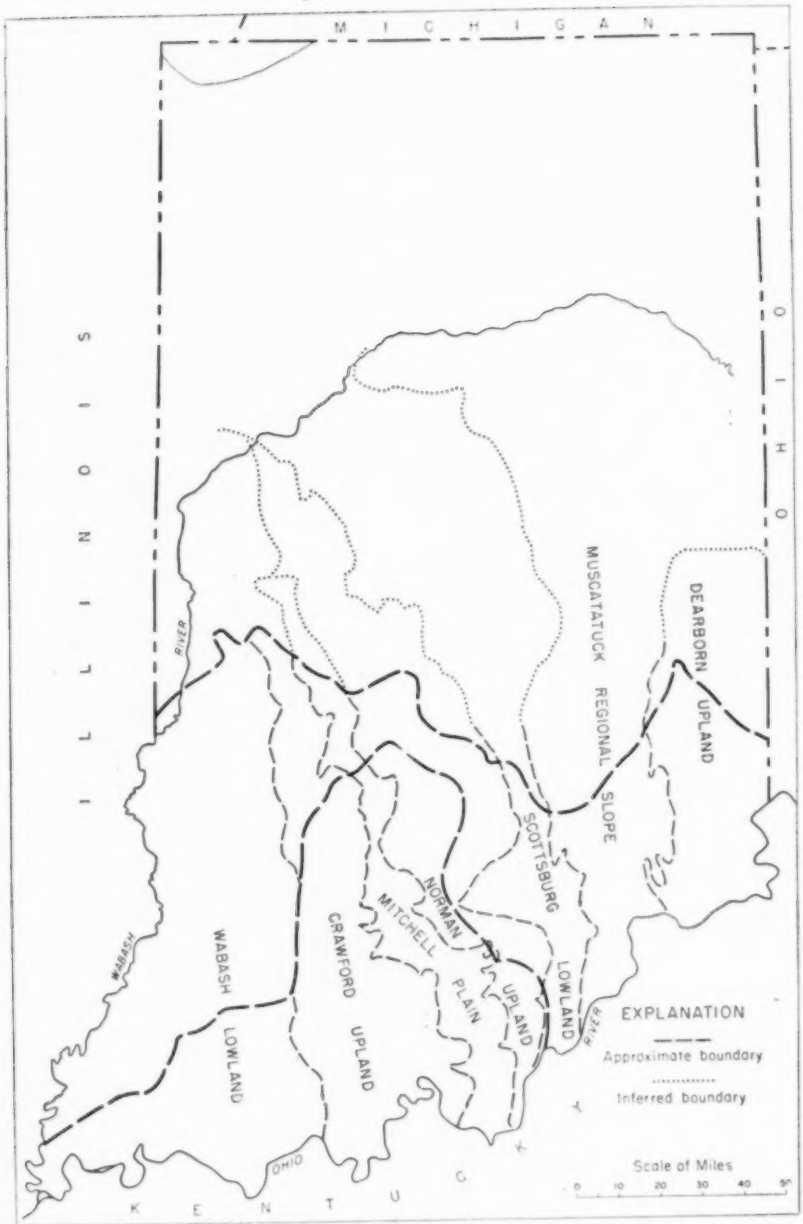

FIG. 3.—Physiographic units of southern Indiana, after Malott (1922), and their approximate boundaries where buried north of the Wisconsin glacial boundary.

Madison. Leverett (1929, p. 53) indicates that the south bluff of the Ohio River is composed of glacial till a short distance above the present mouth of the Kentucky River. On the northwest side, the basin developed on Ordovician rocks is rimmed by resistant Silurian limestones. Streams that flow westward from this scarp enter the Wabash drainage area; those, such as Laughery Creek and Tanner's Creek, which lie on the eastern side of the rim of limestone join the Ohio River above Madison (U.S. Geol. Survey, Aurora, Indiana, quadrangle). The writer thinks it highly unlikely that a small consequent river, such as the preglacial Ohio, could have accomplished the extensive headward erosion and the several piracies necessary to develop a river that crossed the Cincinnati Arch. For this reason, in addition to those listed above, the writer thinks that the preglacial Miami River probably flowed northeastward and that the Ohio River headed on the Muscatatuck regional slope.

WABASH VALLEY

Fidlar (1948, fig. 1) considered the preglacial Wabash River to have headed near Covington in Fountain County. Thornbury (1948, p. 1359) suggested that the Knobstone escarpment formed the preglacial divide in western Indiana between the basins of the Wabash and the Mahomet-Teays. The bedrock drainage pattern which exists in Fountain and Parke counties and the bedrock altitudes in the same area indicate that the divide was more likely about 15 miles south of Covington in northern Parke County and southern Fountain County. South of Parke County, the course of the bedrock Wabash Valley has been described in detail by Malott (1922), Fidlar (1948), Horberg (1950), and others.

LATE TERTIARY EROSION CYCLES

Indiana and the surrounding area were reduced to a gently rolling, old-age, topographic surface during the latter part of the Tertiary, probably in late Miocene or Pliocene time. This surface, the Lexington peneplain, is about 900–1,000 feet in altitude in the Dearborn upland (fig. 3) of southeastern Indiana, and the residual local relief is 100–200 feet. The Lexington peneplain sloped north and west toward the major drainage lines. It had an altitude of 700–800 feet in Wabash County in north-central Indiana (Wayne and Thornbury, 1951, p. 26). The peneplanation probably never was so complete that uplands and lowlands were unrecognizable. On the contrary, some structural and lithologic control of topography must have remained.

A new cycle of erosion called the "Parker" began with the late Tertiary rejuvenation, which caused the streams to entrench themselves 100–200 feet below the level of the peneplain and to develop broad floodplains. Lowlands were developed along the outcrop belts of weaker rocks, particularly those traversed by major streams. Degradation of the Scottsburg lowland of southern Indiana and its buried northward extension began during this period, as did that of the Pennsylvanian lowland (Horberg, 1950, pp. 35–37) of Illinois and Indiana. This cycle was interrupted by glaciation.

The present Scottsburg lowland is a result of degradation throughout the Pleistocene. Inasmuch as a major river system, the Mahomet-Teays, crossed north-central Indiana, the northern end of the lowland should have been eroded more than the southern end during preglacial times. Analysis of drift thickness indicates the existence of a former broad lowland along the outcrop of Devonian and Lower Mississippian shales in central In-

diana, where they are deeply buried beneath Wisconsin glacial deposits. This part of the Scottsburg lowland was formed between the end of the Lexington cycle and the beginning of the Illinoian age, and probably the greater part of the development took place during Parker time. The southern part of the Scottsburg lowland has been further eroded and deepened since the enlargement of the Ohio River following Nebraskan glaciation.

Glaciation caused complications in the Parker cycle, including drainage diversions, static rejuvenation, and alluviation in the major rivers. The present rivers in central Indiana follow largely the courses of Wisconsin or Illinoian sluiceways and in some places show little or no relation to the bedrock topography. Much of the post-Wisconsin drainage is superposed, so that streams now cross buried uplands and valleys with little relationship to either.

In north-central Indiana the main preglacial rivers appear to have been deeply entrenched below broad valleys that were about 100 feet below the level of the Lexington peneplain. The bedrock surface slopes gently toward these valleys, the dissected remnants of which have been interpreted as strath terraces representing the Parker cycle (Wayne and Thornbury, 1951, p. 29). The writer now thinks that these old-age valleys more likely are simply the broad troughs through which rivers flowed across the Lexington peneplain.

DEVELOPMENT OF THE DEEPLY INCISED VALLEYS

Several differences of opinion have arisen concerning the age of the deeply entrenched bedrock valleys. Ver Steeg (1936, pp. 935–936) suggested that the deep erosion along the major valleys ("Deep stage") took place after the disruption of the Mahomet-Teays drainage system and left the Parker strath as a terrace above the entrenched rivers. He states (1936, pp. 936–937): "The writer believes that the deep, comparatively narrow valleys buried beneath the glacial drift of western Ohio and incised below an old surface (Parker) were cut during Deep-stage time. . . . The Deep stage lasted from the flood stage of the Teays to the advent of the Illinoian glacier."

Horberg (1950, p. 70) recognized probable outwash below Aftonian soil in the bottom of the Mahomet Valley in Illinois, thus indicating that the erosion there may have been entirely pre-Pleistocene.

Thornbury (1948, p. 1359) interpreted the deep valley across north-central Indiana as part of the "Deep stage" of the Mahomet-Teays drainage system and suggested that it was eroded before Illinoian glaciation by an interglacial river that headed in western Ohio.

It is entirely possible that the "Deep stage" of the Mahomet-Teays Valley is pre-Pleistocene. Where the Teays disappears beneath glacial drift near Chillicothe, Ohio, its bedrock altitude is 620 feet (Stout and Schaaf, 1931, p. 665). The wide bench above the deep valley in Wabash County, Indiana, which has been interpreted as part of the same valley floor as that in southern Ohio (Wayne and Thornbury, 1951, p. 29) is about 600 feet above sea-level. The length of the valley between Chillicothe and the western edge of Wabash County is at least 210 miles, and the stream course probably was closer to 300 miles long. The middle reaches of this river, therefore, would seem to have had an average gradient of less than 2 inches per mile, a conclusion that is improbable. Differential rates of uplift could have caused such a

gradient, but, if uplift occurred, its effects should be noticeable in steepened gradients of eastward-flowing streams and lessened gradients of other west-flowing streams in Indiana. Such phenomena have not been observed. The deepest part of the Mahomet-Teays Valley in Wabash County, Indiana, is about 410 feet in altitude, or 210 feet lower than at Chillicothe, Ohio, which would provide a maximum average gradient of 12 inches per mile. Horberg (1945, p. 359) calculated the average gradient of the Teays-Mahomet Valley between Chillicothe, Ohio, and Tazewell County, Illinois, as 7 inches per mile; and Stout, Ver Steeg, and Lamb (1943, p. 53) calculated the gradient of the valley across Ohio as 12 inches per mile. For comparison, the average gradient of the present Ohio River from the mouth of the Miami River to its junction with the Wabash River is 4 inches per mile.

It seems rather unlikely that, after ponding and derangement of the Mahomet-Teays drainage system and diversion of the upper part over a divide into the Ohio Valley (fig. 2), the beheaded remains of the Mahomet River somewhere in west-central Ohio could have been capable of entrenching itself deeply below its former level. Instead, one would expect some aggradation by the underfit river after such a loss of volume. More likely, little or no deepening of the Mahomet Valley could have occurred after it was shortened.

The precise glaciation that caused the ponding of the Mahomet-Teays drainage system has not been determined. Leverett (1929, fig. 3), described partly decomposed erratics at various locations in Kentucky. As rafting by icebergs does not satisfactorily account for all the boulders, Leverett (1929, p. 33) postulated an extensive ice sheet of pre-Illinoian age. Thwaites (1947, pl. 3) interprets these boulders as Nebraskan in age.

Diversion of so large a river as the upper part of the Mahomet-Teays into the Ohio Valley most certainly would have caused static rejuvenation to take place and rapid regrading to occur along that valley (Malott, 1922, p. 170). Glaciation of the duration necessary to derange the Mahomet-Teays system permanently could likewise have caused reversal of drainage in the preglacial Miami Valley of southwestern Ohio, which, in cutting in accordance with the deepened Ohio River, would have developed the observed southwesterly gradient of the Miami bedrock valley. A narrow but deep buried trench that is intersected by the present Whitewater River at Connersville, Indiana, may have been formed at the same time by overflow waters impounded in Anderson Valley. Well cuttings show that lacustrine clays more than 160 feet thick lie beneath Wisconsin till in Anderson Valley, but data are insufficient to determine whether this entire thickness represents one or more than one glacial stage. The relatively soft Ordovician shales in the Cincinnati area would have allowed fairly rapid reduction of interstream areas to conform to the new stream orientation. Thus the deep erosion ("Deep stage") of the Ohio River seems to be post-Nebraskan in age.

What appears to be a "valley-in-valley" profile along the Mahomet-Teays in western Ohio and in Wabash County, Indiana, may be a result of only one entrenchment rather than two. The rivers that crossed the gently rolling Lexington peneplain in late Tertiary time undoubtedly flowed through broad valleys and were somewhat below the level of the interstream areas. Following uplift, these rivers became deeply entrenched below their old floodplains.

The steep sides that occur along the bedrock valleys in east-central Indiana probably result from the greater resistance to erosion of the Niagaran limestones of the region as compared with shales both east and west of that part of the Cincinnati and Wabash arches. The valley floor had reached a stage of early maturity. Strath terraces have not been found along the upper reaches of the Anderson Valley (fig. 1). In its lower reaches, this valley passes through Devonian and Mississippian shales that underlie the Scottsburg lowland, and the river was able to erode a much wider valley than in limestone.

Horberg (1945, p. 354) drew generalized contours on the Sangamon interglacial surface of weathering in central Illinois. On the basis of this map and on the lithology of well cuttings along the Mahomet Valley, he concluded that the Mahomet River was insignificant as a drainage line across Illinois after Kansan glaciation. Outwash gravels and sands are extensive along the Mahomet Valley in Indiana and indicate that it was a sluiceway until late Tazewell time in this state. A possible explanation for the presence of abundant outwash in Indiana and its absence in Illinois is suggested by the deep erosion ("Deep stage") along the Wabash Valley that seems to extend north to the Mahomet Valley. At one time Fidlar (1943, p. 415) thought that the deep Mahomet-Teays Valley across Indiana was a continuation of the Wabash. Certainly the divide between the two drainage systems must have been low. Kansan ice may have impounded water in the Mahomet Valley above Warren County, Indiana, and diverted the drainage of the upper part of the basin into the Wabash Valley (fig. 4). A

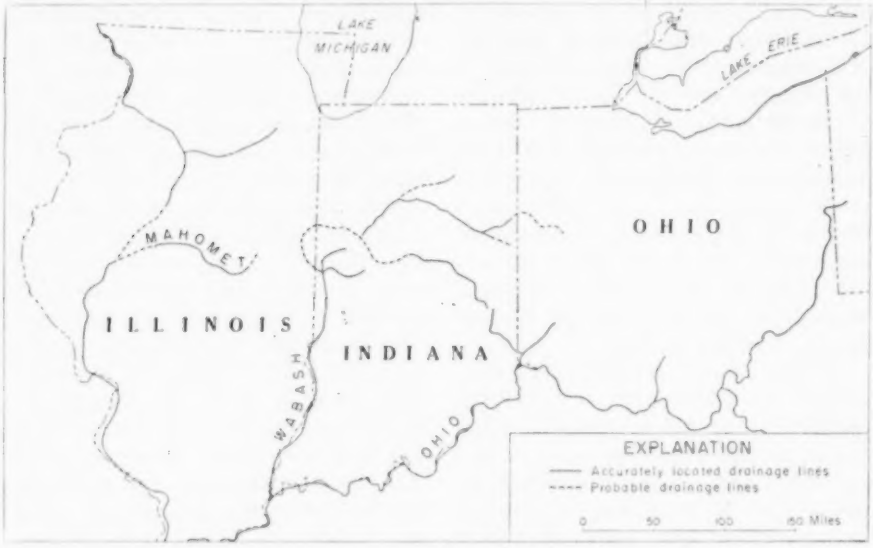

FIG. 4.—Drainage modifications that resulted from Kansas and Illinoian glaciations. Ohio data after Stout, Ver Steeg, and Lamb (1943); Illinois data after Horberg (1950).

buried valley that contains considerable outwash crosses the bedrock valley of the Wabash in northern Vermillion County. This valley is relatively narrow and is a continuation of buried Danville Valley in Illinois (Horberg, 1951, p. 71). Static rejuvenation would have caused the overfit river to begin cutting its course to grade immediately. Such a history could explain the peculiarity of the deeply entrenched bedrock valley that appears to head so close to a major preglacial river. Presence (reported by well-drillers) of silt and fine sand in the Mahomet Valley in Tippecanoe, White, and Carroll counties, Indiana, and exposures of thick outwash deposits overlain by till in Fountain and Warren counties, Indiana, support this interpretation, but data are admittedly inconclusive. Much of the "Deep stage" of the Wabash Valley, according to this interpretation, would have been cut during Kansan and Yarmouth time as a result of an increase in volume from diversion of part of the Mahomet River.

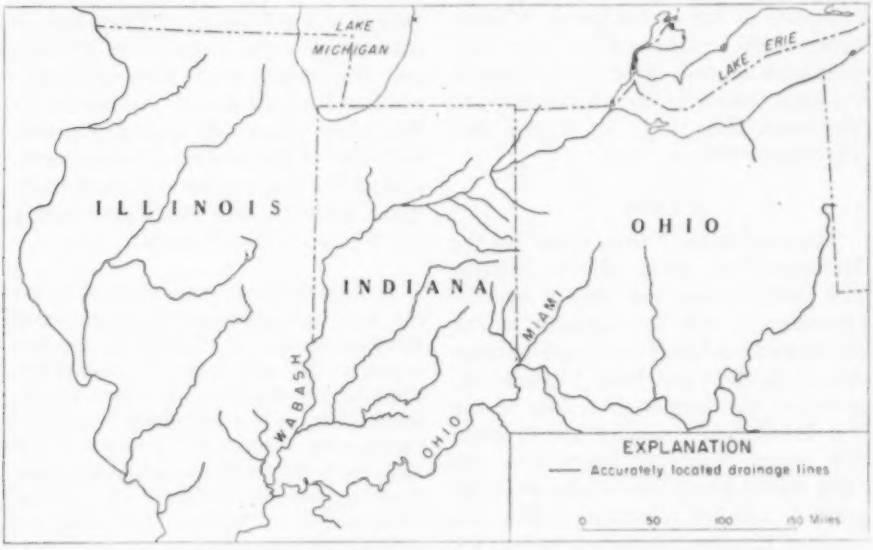

FIG. 5.—Principal post-Wisconsin drainage systems of Indiana and adjoining states

LATE PLEISTOCENE VALLEY-FILLING

The extent of Illinoian glaciation in Indiana has been discussed thoroughly by Leverett and Taylor (1915), Malott (1922), Thornbury (1937), and others. Some of the more important events that took place during this glacial age were the beginning of extensive deposition along the Ohio River and many of its tributaries, and continued filling of the Mahomet and its tributaries with till and glaciofluviatile deposits. As much as 160 feet of calcareous lacustrine clay was deposited in Anderson Valley, most of it probably during Illinoian time. Malott (1922, pp. 140–141) suggested that the great volume of Illinoian and Wisconsin outwash enlarged the Mississippi delta and thereby forced the Mississippi River

and its tributaries to aggrade their valleys. This may have been a contributing factor, along with the rising sea-level during deglaciation (Turnbull, Krinitzsky, and Johnson, in: Trask, 1950, p. 210).

Early Wisconsin (Tazewell) ice provided great quantities of outwash, which aggraded major sluiceways, such as the Ohio and Wabash rivers (Fidlar, 1948, pl. 3). The upper Wabash apparently still followed the course of the ancient Mahomet as far as Lafayette. A readvance of the ice covered the thick outwash with till, and, when the ice melted, the upper course of the present Wabash was established (fig. 5; Wayne and Thornbury, 1951, p. 32).

SUMMARY

The deep valley ("Deep stage") of the Mahomet-Teays across Illinois, Indiana, and western Ohio was eroded contemporaneously with the surfaces in the abandoned reaches of the Teays drainage system in Ohio and West Virginia and probably was preglacial. A lobe of the Nebraskan glacier ponded and diverted the upper Mahomet-Teays into the Ohio Basin, the Minford silts were deposited, and the direction of flow was reversed in some tributaries, such as the Miami. Erosion of the "Deep stage" of the Ohio Valley next occurred in post-Nebraskan time (Aftonian, possibly longer), when the Mahomet became an underfit stream that headed in western Ohio. Later the upper Mahomet probably was diverted into the Wabash Valley as a result of Kansan glaciation, and the "Deep stage" was cut along the lower Wabash in Yarmouth time. Extensive valley-filling in all the "Deep-stage" valleys of the Mississippi River and its tributaries began during the Illinoian age. The remains of the Mahomet drainage line in Indiana was obliterated by Wisconsin (Tazewell) glacial deposits, and many of the present valleys in north-central Indiana were eroded along meltwater drainage lines established during the Wisconsin glacial retreat.

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THE 10 PER CENT $CaAl_2Si_2O_8$ PLANE IN THE SYSTEM $CaSiO_3$-$Ca_2Al_2SiO_7$-$NaAlSiO_4$-$CaAl_2Si_2O_8$[1]

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ABSTRACT

The high-temperature equilibrium relationships in the 10 per cent $CaAl_2Si_2O_8$ plane in the $CaSiO_3$-$Ca_2Al_2SiO_7$-$NaAlSiO_4$-$CaAl_2Si_2O_8$ system have been investigated by the quenching method. The results contribute to the understanding of the origin and differentiation of alkaline rocks.

INTRODUCTION

A number of investigations have been carried out on portions of the four-component system Na_2O-CaO-Al_2O_3-SiO_2. Albite, anorthite, nepheline, gehlenite (melilite), and wollastonite are mineral compounds in this system, which is, therefore, at the very heart of the problem of the possible differentiates from a melilite-nephelinite magma. Determined data are sufficient to fix all the general relations within that portion of the quaternary system of immediate interest and to permit description of the courses of crystallization within it in a qualitative manner (Bowen, 1945). The present plane is but one of the many requiring investigation in order to realize fully the quantitative aspects of the Na_2O-CaO-Al_2O_3-SiO_2 system. The position of the plane in the subsidiary tetrahedron Wol-Geh-Ne-An is indicated in figure 1. The four principal faces of this tetrahedron have already been investigated: Wol-Geh-An, Rankin and Wright (1915); Wol-Ne-An, Gummer (1943); Wol-Geh-Ne, Juan (1946, published 1950); Geh-Ne-An, Goldsmith (1947). The present investigation is the first step in a plan to examine planes parallel to the Wol-Geh-Ne face at intervals of 10 per cent anorthite.

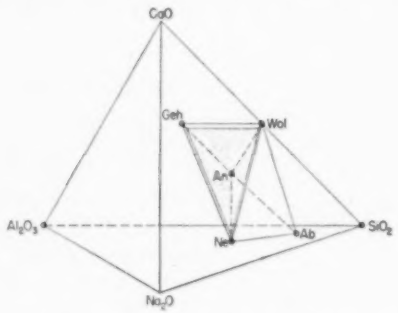

FIG. 1.—Tetrahedral diagram of the quaternary system Na_2O-CaO-Al_2O_3-SiO_2 showing the position of the 10 per cent anorthite plane (stippled area) in the subsidiary tetrahedron wollastonite (Wol)-gehlenite (Geh)-nepheline (Ne)-anorthite (An). The subsidiary tetrahedron nepheline-wollastonite-anorthite-albite (Ab) has important bearing on the crystallization of liquids originating in the first-named subsidiary tetrahedron. Mol per cent.

Forty mixtures were prepared and subjected to thermal study by means of the quenching method developed in the Geophysical Laboratory. Temperatures were measured with a platinum–platinum 90 rhodium 10 thermoelement calibrated at the melting points of lithium metasilicate (1,201° C.) and diopside (1,391°.5 C.). The products of the quenching runs were examined microscopically to determine the phases present at equilibrium. The equilibrium relations at the liquidus are given in figure 3.

[1] Manuscript received November 12, 1951. The work on which this paper is based was done at the University of Chicago in the spring of 1946, before more recent papers bearing on the subject were published.

PROPERTIES OF COMPOUNDS

PSEUDOWOLLASTONITE

The high-temperature form of $CaSiO_3$ is monoclinic, and the indices of refraction, according to Ferguson and Merwin (1919), are $\alpha = 1.610$, $\beta = 1.611$, $\gamma = 1.654$. The crystals commonly appeared as hexagonal plates exhibiting low birefringence. When seen on edge, the plates appeared as laths with negative elongation and high birefringence.

CARNEGIEITE

The high-temperature modification of $NaAlSiO_4$ is isometric at the temperature of formation and inverts at approximately 690° C. during the quenching operation. The indices of refraction of the modification at room temperature are given by Bowen (1912) as $\alpha = 1.509$, $\gamma = 1.514$. In the present investigation the crystals appeared as rounded globules, frequently arranged in chains or clusters. Intricate twinning lamellae were observed universally.

NEPHELINE

Pure carnegieite inverts to nepheline, the low form of $NaAlSiO_4$, at 1,254° C. (Greig and Barth, 1938). In the compositions under investigation the inversion takes place from 1,293° to 1,321° C. as a result of solid solution. Nepheline appears as hexagonal prisms, commonly minute. When elongation is determinable, it is, in most cases, positive, but it may be either positive or negative in crystals of the same run. Extremely low birefringence was observed. The indices of refraction of pure nepheline are given by Bowen (1912) as $\omega = 1.537$ and $\epsilon = 1.533$; however, solid solution of lime compounds may raise the indices of refraction to $\omega = 1.548$ and $\epsilon > 1.548$ (Goldsmith, 1947; Smalley, 1947). The nepheline in those few runs which were amenable to optical study gave indices of refraction in the range $\omega = 1.537$–1.543 and $\epsilon' = 1.533$–1.541.

GEHLENITE

One of the end-members of the melilites is gehlenite. It is tetragonal with $\omega = 1.669$ and $\epsilon = 1.658$. The melilites formed in the present compositions have indices of refraction, as well as birefringence, considerably lower than those for pure gehlenite. Although prismatic crystals were most common, other forms were found. For those compositions in the melilite field whose liquidus temperature was above 1,300° C., cubes and, in one case, parallelopipeds were observed; below 1,300°, square, rectangular, or octagonal plates were observed. The melilite occurs as rounded globules, similar to those of carnegieite, in the vicinity of the boundary curve between the fields of melilite and nepheline solid solutions near the 90 $Ca_2Al_2SiO_7$ 10 $CaAl_2Si_2O_8$—90 $NaAlSiO_4$ 10 $CaAl_2Si_2O_8$ join.

THERMAL DATA

The data for the join

90 $CaSiO_3$ 10 $CaAl_2Si_2O_8$—90 $Ca_2Al_2SiO_7$ 10 $CaAl_2Si_2O_8$

were obtained from the work of Rankin and Wright (1915). The point 90 $CaSiO_3$ 10 $CaAl_2Si_2O_8$ is from Osborn and Schairer (1941), the point 90 $Ca_2Al_2SiO_7$ 10 $CaAl_2Si_2O_8$ is from Osborn (unpublished data), and the point 90 $NaAlSiO_4$ 10 $CaAl_2Si_2O_8$ is from Bowen (1912). The data for the join

90 $CaSiO_3$ 10 $CaAl_2Si_2O_8$—90 $NaAlSiO_4$ 10 $CaAl_2Si_2O_8$

are from Gummer (1943). The significant thermal data necessary for fixing the equilibrium relationships in the 10 per cent $CaAl_2Si_2O_8$ plane in the system

$CaSiO_3$-$Ca_2Al_2SiO_7$-$NaAlSiO_4$-$CaAl_2Si_2O_8$

are given in table 1. The phase relations are shown in figures 2 and 3. The isofract diagram given in figure 4 is based on the homogeneous glasses prepared in the plane.

The boundary curves divide the plane into four fields, namely, those of pseudowollastonite, melilite, nepheline solid solutions, and carnegieite solid solutions. Crystallization in the fields must be considered in terms of the quaternary Na_2O-CaO-Al_2O_3-SiO_2. The liquidus surfaces have essentially similar slopes and meet at the piercing point. The piercing point is on the curve which represents the common boundary of the three-dimensional fields of pseudowollastonite, nepheline solid solutions, and melilite in the quaternary system.

The temperature of the piercing point must be between 1,169° and 1,178° C., respectively, the temperature given by Gummer (1943, p. 515) and that determined from the data of Juan (1950, pp. 7–8). An examination of the runs for the three mixes nearest the piercing point suggests that the point ought to be higher than 1,163° and lower than 1,176°. These runs only set the limits because of the nonternary nature of the system. It is believed, therefore, that 1,173° is the best estimate of the temperature of the piercing point. If the temperature gradient were precisely known along the boundary curve, it would be possible to fix the composition of the liquid from which pseudowollastonite, melilite, and nepheline solid solutions could crystallize simultaneously.

The maximum which was found in the nepheline solid solution–pseudowollastonite boundary curve in the systems investigated by Juan and by Gummer does not appear in this plane; the temperature falls continuously from the 90 $CaSiO_3$ 10 $CaAl_2Si_2O_8$—90 $NaAlSiO_4$ 10 $CaAl_2Si_2O_8$ join to the piercing point.

NATURE OF SOLID SOLUTIONS

The univariant curve between the fields of nepheline and carnegieite is not isothermal. The discontinuity on the liquidus caused by the inversion occurs many degrees higher than the inversion temperature of the pure compound. The increase and the large range of temperature are the result of solid solution in the nepheline or solid solution in both nepheline and carnegieite, provided that it is greater in the nepheline. It is reported that the "molecules" $CaSiO_3$ (Foster, 1942, p. 160), $Ca_2Al_2SiO_7$ (Smalley, 1947, p. 35), $CaAl_2Si_2O_8$ (Bowen, 1912, p. 559), and $CaAlAlO_4$ (Goldsmith, 1949, p. 488) can enter nepheline in solid solution.[2] The few indices of refraction for nepheline which were measured are given in table 2.

The indices of refraction of the melilites crystallizing in this system are considerably lower than those for pure gehlenite (table 3). In a general way the indices of refraction increase with the content of gehlenite in the initial mix. At first thought this might be regarded as indicating that both $CaSiO_3$ and $NaAlSiO_4$ enter into solid solution with gehlenite; it has been shown, however, that the solid solution is of a quaternary nature involving the "molecules" $Na_2SiSi_2O_7$ and possibly $CaAl_2Al_2O_7$ (Goldsmith, 1948).

[2] It is convenient in phase-equilibrium studies to think of solid solutions in terms of "molecules," even though the solid solutions take place structurally in terms of ions.

TABLE 1

THERMAL DATA FOR THE 10 PER CENT $CaAl_2Si_2O_8$ PLANE IN THE SYSTEM $CaSiO_3$-$Ca_2Al_2SiO_7$-$NaAlSiO_4$-$CaAl_2Si_2O_8$

Mix No.	Weight Per Cent*				Initial Condition	Time (Min.)	Temp. (° C.)	Final Condition†
	An	Pwol	Geh	Ne				
	Primary-Phase Pseudowollastonite							
4	10.0	37.8	1.8	50.4	Glass	45	1,177	Glass
					Glass	45	1,175	Pwol, glass
					Glass	60	1,171	Pwol, nes, glass
					Glass	60	1,168	Pwol, nes, glass
9	10.0	44.1	5.4	40.5	Glass	30	1,219	Glass
					Glass	30	1,216	Pwol, glass
18	10.0	46.8	11.7	31.5	Glass	30	1,237	Glass
					Glass	30	1,233	Pwol, glass
20	10.0	49.5	17.1	23.4	Glass	30	1,254	Glass
					Glass	30	1,250	Pwol, glass
22	10.0	54.0	27.0	9.0	Glass	30	1,287	Glass
					Glass	30	1,283	Pwol, glass
30	10.0	58.5	9.0	22.5	Glass	35	1,279	Mel, pwol, glass
					Glass	30	1,307	Glass
31	10.0	63.0	18.0	9.0	Glass	30	1,305	Pwol, glass
					Glass	30	1,344	Glass
					Glass	30	1,340	Pwol, glass
	Primary-Phase Melilite							
1	10.0	40.5	4.5	45.0	Glass	45	1,195	Glass
					Glass	45	1,191	Mel, pwol, glass
					Glass	45	1,179	Mel, pwol, glass
2	10.0	36.0	9.0	45.0	Glass	30	1,211	Glass
					Glass	35	1,206	Mel, glass
3	10.0	31.5	9.0	49.5	Glass	30	1,218	Glass
					Glass	45	1,213	Mel, glass
5	10.0	36.0	3.6	50.4	Glass	30	1,183	Glass
					Glass	30	1,180	Mel, glass
					Glass	30	1,175	Mel, glass
					Glass	30	1,166	Mel, nes, glass
					Glass	15	1,163	Mel, nes, pwol, glass
8	10.0	23.4	13.5	53.1	Glass	30	1,239	Glass
					Glass	30	1,234	Mel, nes, glass
10	10.0	13.5	20.7	55.8	Glass	30	1,266	Glass
					Glass	30	1,262	Mel, glass
					Devit	30	1,262	Mel, glass
11	10.0	6.3	27.0	56.7	Glass	30	1,277	Glass
					Devit	30	1,275	Mel, nes, glass
					Glass	30	1,275	Mel, nes, glass
13	10.0	10.8	24.3	54.9	Glass	30	1,275	Glass
					Glass	30	1,271	Mel, glass
					Glass	15	1,262	Mel, nes, glass
					Devit	30	1,254	Mel, nes, glass
24	10.0	49.5	23.5	18.0	Glass	30	1,262	Glass
					Glass	30	1,258	Mel, glass
26	10.0	44.1	9.9	36.0	Devit	40	1,228	Glass
					Devit	40	1,224	Mel, pwol, glass
32	10.0	36.0	18.0	36.0	Glass	30	1,260	Glass
					Glass	30	1,256	Mel, glass

* An = anorthite; Pwol = pseudowollastonite; Geh = gehlenite; Ne = nepheline.
† Mel = melilite; Nes = nepheline solid solutions; Cgs = carnegieite solid solutions.

TABLE 1—*Continued*

Mix No.	Weight Per Cent*				Initial Condition	Time (Min.)	Temp. (° C.)	Final Condition†
	An	Pwol	Geh	Ne				
	Primary-Phase Melilite—*Continued*							
33	10.0	40.5	27.0	22.5	Glass	30	1,308	Glass
					Glass	30	1,304	Mel, glass
34	10.0	22.5	27.0	40.5	Glass	30	1,312	Glass
					Glass	30	1,308	Mel, glass
35	10.0	27.0	36.0	27.0	Glass	30	1,366	Glass
					Glass	30	1,362	Mel, glass
36	10.0	40.5	40.5	9.0	Glass	30	1,404	Glass
					Glass	30	1,400	Mel, glass
37	10.0	9.0	40.5	40.5	Glass	30	1,362	Glass
					Glass	30	1,358	Mel, glass
	Primary-Phase Carnegieite Solid Solutions							
21	10.0	4.5	11.7	73.8	Glass	40	1,358	Glass
					Glass	45	1,354	Cgs, glass
23	10.0	0.0	18.9	71.1	Glass	50	1,350	Glass
					Glass	30	1,346	Cgs, glass
25	10.0	10.8	7.2	72.0	Glass	40	1,321	Glass
					Glass	40	1,317	Cgs, glass
27	10.0	4.5	16.2	69.3	Glass	30	1,327	Glass
					Glass	30	1,323	Cgs, glass
29	10.0	0.0	21.6	68.4	Glass	30	1,333	Glass
					Glass	30	1,329	Cgs, glass
	Primary-Phase Nepheline Solid Solutions							
6	10.0	36.0	1.8	52.2	Glass	30	1,186	Glass
					Glass	30	1,181	Nes, glass
7	10.0	31.5	6.3	52.2	Glass	30	1,204	Glass
					Glass	30	1,201	Nes, glass
12	10.0	0.0	33.3	56.7	Glass	30	1,290	Glass
					Glass	30	1,287	Nes, glass
					Devit	30	1,287	Nes, glass
					Glass	30	1,283	Mel, nes, glass
14	10.0	0.0	28.8	61.2	Glass	60	1,299	Glass
					Glass	60	1,297	Nes, glass
15	10.0	13.5	8.1	68.4	Glass	30	1,295	Glass
					Glass	30	1,292	Nes, glass
16	10.0	6.3	18.0	65.7	Glass	30	1,309	Glass
					Glass	30	1,306	Nes, glass
17	10.0	9.0	11.7	69.3	Glass	30	1,312	Glass
					Glass	30	1,309	Nes, glass
19	10.0	0.0	24.3	65.7	Glass	30	1,316	Glass
					Glass	30	1,312	Nes, glass
28	10.0	9.9	22.5	57.6	Glass	30	1,273	Glass
					Glass	30	1,270	Nes, glass
39	10.0	27.0	4.5	58.5	Glass	30	1,229	Glass
					Glass	30	1,225	Nes, glass
40	10.0	18.0	9.0	63.0	Glass	30	1,266	Glass
					Glass	30	1,262	Nes, glass

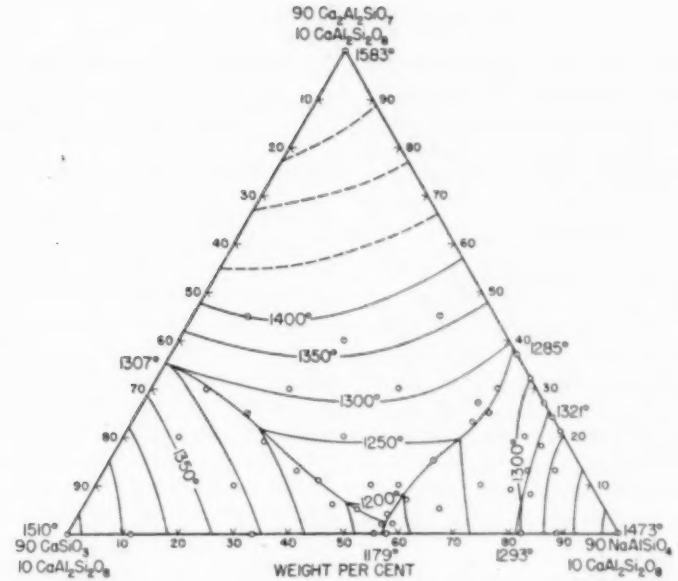

FIG. 2.—Equilibrium diagram for the 10 per cent $CaAl_2Si_2O_8$ plane in the system $CaSiO_3$—$Ca_2Al_2SiO_7$—$NaAlSiO_4$—$CaAl_2Si_2O_8$, with isotherms.

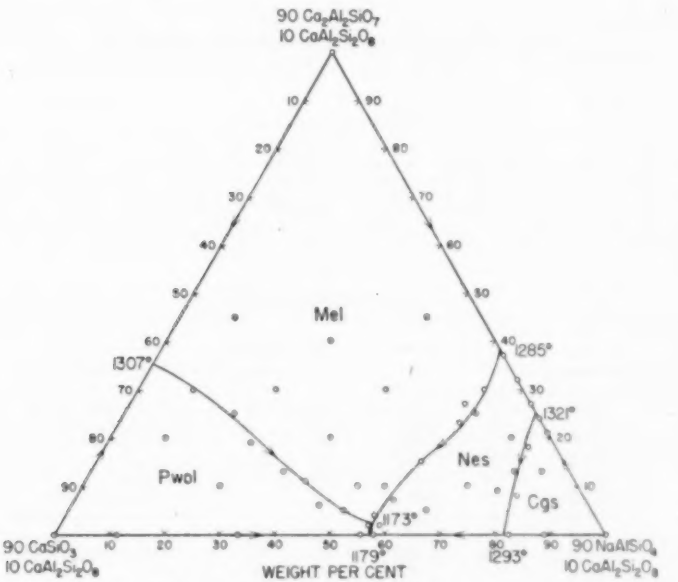

FIG. 3.—Equilibrium diagram for the 10 per cent $CaAl_2Si_2O_8$ in the system $CaSiO_3$—$Ca_2Al_2SiO_7$—$NaAlSiO_4$—$CaAl_2Si_2O_8$, with isotherms omitted.

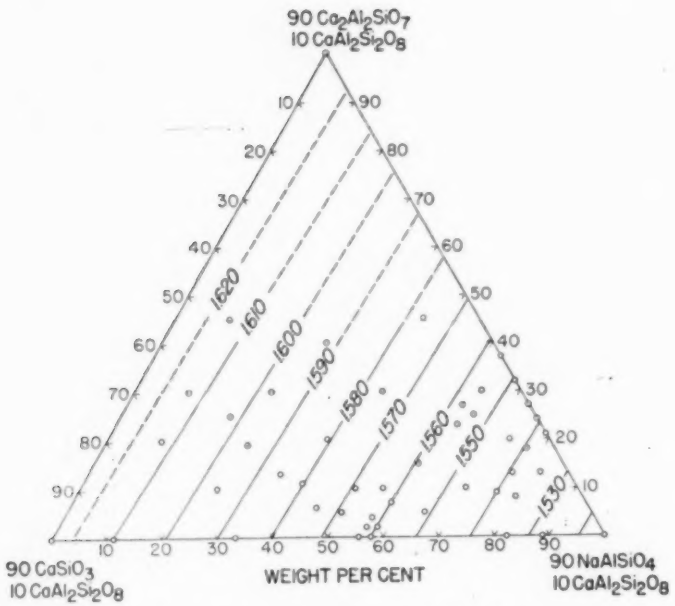

FIG. 4.—Isofract diagram for glasses in the 10 per cent $CaAl_2Si_2O_8$ plane in the system $CaSiO_3$—$Ca_2Al_2SiO_7$—$NaAlSiO_4$—$CaAl_2Si_2O_8$.

TABLE 2

OPTICAL PROPERTIES OF NEPHELINE SOLID SOLUTIONS

Mix No.	Per Cent Nepheline	Temp. (° C.)	ω	ϵ'	Elongation
16	65.7	1,296	1.543	1.541	−(+)
28	57.6	1,266	1.538	1.535	+
39	58.5	1,225	1.537	1.533	+(−)
40	63.0	1,262	1.538	1.534	+

TABLE 3

OPTICAL PROPERTIES OF MELILITE

Mix No.	Per Cent Gehlenite	Temp. (° C.)	ω	ϵ'	Elongation	Form
2	9.0	1,201	1.649	1.646	+	Cubes, square plates
3	9.0	1,200	1.647	1.645	+	Octagonal plates
11	27.0	1,271	1.653	1.651	+	Rectangular plates, globules
12	33.3	1,283	1.655	1.653	N.O.*	Globules
13	24.3	1,254	1.651	1.649	N.O.	Globules
22	27.0	1,279	1.654	1.651	+	Rectangular plates
26	9.9	1,216	1.647	1.645	+	N.O. (partially melted)
33	27.0	1,287	1.655	1.652	N.O.	Cubes
34	27.0	1,304	1.655	1.653	N.O.	Cubes, few square plates
35	36.0	1,358	1.657	1.654	N.O.	Cubes
36	40.5	1,370	1.658	1.656	N.O.	Cubes
37	40.5	1,358	1.658	1.656	+	Parallelopipeds

* N.O. = not observed.

COURSES OF CRYSTALLIZATION

All mixtures in the 10 per cent $CaAl_2Si_2O_8$ plane in the system

$CaSiO_3$-$Ca_2Al_2SiO_7$-$NaAlSiO_4$-$CaAl_2Si_2O_8$

are quaternary. In the pseudowollastonite field the composition of the liquid initially changes directly away from the $CaSiO_3$ apex of the tetrahedron and out of the plane under investigation. This is based on the probability that pseudowollastonite crystallizes out as a pure compound. In the fields of melilite and nepheline solid solutions the composition of the liquid during crystallization can be found only by further experiment. Discussion of the complex courses of crystallization must await the study of all planes within the fundamental quaternary system; however, the general relations have already been presented in a previous paper (Bowen, 1945). There it was shown that on fractional crystallization the liquid would yield products grading from melilite nephelinites through simplified tephrites to phonolites. If equilibrium obtains, the final products from liquids in the investigated plane would be melilite, nepheline solid solutions, and pseudowollastonite.

CONCLUSION

The understanding of igneous rocks has been greatly enlarged by phase-equilibrium studies. As more and more components are studied, a closer approach to the natural magma is achieved. The present work is but one of the many steps required for a full understanding of the system Na_2O-CaO-Al_2O_3-SiO_2 in which the compositions of the bulk of the alkaline rocks can be represented.

ACKNOWLEDGMENTS.—The writer is indebted to the faculty of the Department of Geology of the University of Chicago for the laboratory facilities which made this research possible. He is particularly grateful to Dr. Norman L. Bowen for his guidance throughout the study and for his kindness in reviewing the manuscript.

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REVIEWS

Geography in the Twentieth Century. Edited by GRIFFITH TAYLOR. New York: Philosophical Library, 1951. Pp. x+630; figs. 56. $8.75.

Prepared as a "study of the growth, fields, techniques, aims, and trends of geography," this volume represents the joint efforts of twenty authors. It is at once a careful sampling and a skilfully edited summary of geographical thought during the last half-century and, as such, is a unique and valuable contribution. More than six hundred pages of data and sober discussion could well be pretty deadly stuff, but the potential reader need not be frightened away from this book by its somewhat encyclopedic aspect, even though he will quickly recognize most of the chapters as both comprehensive and authoritative. Not only is the general approach philosophical rather than merely descriptive, but much of the writing itself is lucid and pleasantly tidy. Many provocative ideas and arguments, presented in a wide variety of forms, are almost certain to raise questions in the reader's mind, particularly if he is an engineer or scientist with normal interest in the techniques of converting data into sound answers. He may react favorably, or he may react unfavorably, to some of the arguments, but he assuredly will react!

The book comprises three sections, the first of which deals with the development of geography as a science, together with its philosophical basis and the several schools of thought that evolved at various times in different parts of the world. The introductory chapter, provided by Griffith Taylor, the editor, is extraordinarily broad in scope and sets a strong pace for the rest of the book. Among its most attractive features is a succinct discussion of the old but not-yet-dead "theocratic" view of geography, in which the well-being of man is held to be the basic aim of an omniscient providence; the later and now-very-active "geocratic" or environmental view, which places emphasis upon physical factors as the chief control of human activities and development; and the "we-ocratic," or possibilist view, in which man is favored as master of his own geographic destinies and the natural landscape is subordinated to the "cultural landscape." These points of view are later treated in detail by George Tatham, who also contributes a historical chapter on geography in the nineteenth century. Other chapters summarize the development of geography in France, Germany, Czechoslovakia, and Poland. Interestingly enough, an attempt was made by the editor to obtain contributions from Russian and Jugoslav scientists, but no satisfactory contact could be made.

The second section of the book comprises four chapters that are concerned with geomorphology, meteorology, climate, and soils and six chapters that present regional discussions in the light of these environmental factors. These 227 pages thus deal with the most fundamental elements of geography and, as such, constitute the fulcrum of the book. The third section is concerned with aviation; field work; political, social, and racial problems; and other more general topics of broad application. In many respects, such chapters as "Geography and Empire," "Racial Geography," "Geography and Aviation," and "Geopolitics and Geopacifics" are the most stimulating in the book, but they will not yield a full return to the reader who skips over the material in Parts I and II. A good summary of all three parts is presented by the editor in the introductory chapter (pp. 19–27).

The selection of contributing authors was plainly a careful one, and such well-known authorities as S. van Valkenburg, S. W. Wooldridge, Isaiah Bowman, L. Dudley Stamp, Charles B. Fawcett, and Ellsworth Huntington are represented by excellent chapters. It should be added that several of the other chapters, the work of younger men, are fully as good and include some of the most vigorous writing to be found in the book. Each contributor uses his own analytical approach and mode of presentation, which is at once an advantage and a disadvantage to the reader. Although the over-all balance of treatment is thereby improved, an essentially segmented coverage of so broad a field as geography inevitably leaves sutures and gaps in the final product; these, however, have been held to a minimum by the editor, who has integrated the numerous contributions with considerable success.

Scientists in other fields may be somewhat less than favorably impressed by many of the

attempts at quantitative treatment of data and concepts, and perhaps especially by the fundamentally subjective approach to some problems. The geographer's fondness for maps, diagrams, and charts is well shown in several chapters, but the intrinsic accuracy of these helpful illustrations rarely is indicated or even suggested. One wonders, for example, about curves that are based upon three control points and about simple block diagrams that show the relations of several highly complex and imperfectly known variables. One might question, too, the derivations of several formulas on the basis of obviously incomplete data. The geologist will be shocked at some of the overabbreviated, misleading, or even incorrect definitions of terms in the "concise glossary of geographical terms" that concludes the volume.

Another question that most readers will ask again and again relates to possible "sampling errors" in most of the geographic analyses. This problem is undoubtedly recognized by many geographers, and John Kerr Rose, in his discussion of geography in the federal government, states (pp. 575–576): "Field methods, particularly elementary techniques of survey, which do not adequately consider the problems of representativeness of samples, are of no particular help—a good course in elementary statistical methods would be used many times more frequently." The somewhat "personalized" techniques employed by at least four of the authors raises some doubt in the reader's mind as to just where scientific analysis ends and authoritarianism begins. These men—and the editor himself appears to be one of them—repeatedly call attention to their own views in apparent attempts to increase the weight of their arguments.

Despite these relatively minor shortcomings, this book is good reading. In particular, the investigator who deals with the more "exact" areas of science and engineering will better his understanding of geography, whose place in science is not easy to determine. In its dominantly physical or "environmental" facets it is closely related to such well-established fields as astronomy, botany, geology, and physics. In contrast, those facets that form the human side of geography are related to anthropology, economics, history (including politics), and sociology.

Evidently the geographer himself is at times a bit bewildered by the internal complexity of his field, and certainly the attempts of a few geographers to analyze environmental data in rigorous fashion have been blocked again and again by the "human" aspects of the problems. "Human" geography, or the "ontography" of the late William Morris Davis, characteristically involves data that are elusive, inconstant, and all too often misleading. It is evident that such data are not susceptible to the rigorous treatment ordinarily demanded, for example, by the chemist or physicist, and they introduce maddening uncertainties into most geographic problems. And even in those problems that are perfectly free from ontographic complications, the geographer finds his "environmental" data none too promising. Here he joins the physical geologist, who must lean heavily—and perhaps shakily—upon the doctrine of uniformitarianism, who must be willing to make use of semiquantitative data (even though he is not necessarily content with these data), and who must extrapolate far beyond his area of observed relations.

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Optical Crystallography. By ERNEST E. WAHLSTROM. 2d ed. New York: John Wiley & Sons, Inc., 1951. Pp. viii+247; figs. 217. $4.50.

The new edition is very similar to that of 1943 (see this *Journal*, 1945, vol. 53, p. 213) except for (1) the addition of an appendix (only 8 pp.; too short to be of value) on the universal stage; (2) an over-all increase of 41 pages and 11 figures; and (3) a notable increase in stippled perspective drawings. The latter are, in the main, done exceptionally well. The book, considerably improved in detail from the original edition, is recommended highly for a beginning text on a qualitative level. After such praise it is perhaps not unfair to list a few adverse criticisms.

Figures 2 and 4 (pp. 95, 97) would be much more valuable if figures 3 and 5, respectively, were directly above them. Figures 8–12 and 14 (pp. 77–81) are based on Huygens' construction; in a book in which the indicatrix is featured, why not use it here? When the tube is raised (fig. 3*B*, p. 50) the image goes down, not up. Figure 2 (p. 49) violates a first principle of optics; converging rays *1* and *2* entering a plane-parallel–surfaced crystal plate cannot be parallel on leaving it. The numbers on the dots in the third diagram of figure 13*A* (p. 80) are reversed. The "real image" actually seen when the cono-

scope is used (fig. 1, p. 113) is at the cross-hairs, not at the Bertrand lens. The denominators in the ellipse equation (p. 111) should be reversed and squared. It is to be regretted that $\alpha, \beta, \gamma, \epsilon$, and ω have been dropped in favor of more cumbersome designations. Where is the confusion if α (alone) refers to a direction of vibration, or (if in association with a number) an index of refraction, or (if with an angle) an interaxial angle? The context never permits any doubt. Properly speaking, there is no such thing as a chemical or petrographic (mineralogic, crystallographic, geologic) or biologic microscope (p. 93); some of the scientists in all these fields use a polarizing microscope.

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The System of Mineralogy (of the Danas), Vol. 2. By CHARLES PALACHE, HARRY BERMAN, and CLIFFORD FRONDEL. 7th ed. New York: John Wiley & Sons, Inc., 1951. Pp. xi+1124; figs. 501. $15.00.

This volume, one-third larger than the first, completes this monumental revision, except for silica and the silicates and certain organic materials (hydrocarbons, etc.). The halides, carbonates, sulfates, and phosphates with allied compounds are described in terms of 41 classes, 95 types, and 663 numbered species. One class is also devoted to 8 organic minerals of 2 types. There are 237 phosphates, 149 sulfates, 83 halides, 68 carbonates, 45 borates, and not over 20 in any other family.

The general style, quality, and coverage are familiar from the first volume. This volume is almost entirely description; only about the equivalent of fourteen pages is devoted to introductory material regarding crystallochemical relationships within any one family, class, type, or series or between certain of these categories. The general quality of the work is exceedingly high; this holds for both authors and publishers. The praise lavished on the first volume (Fisher, 1946) belongs here as well. The work is truly indispensable for the mineralogist; it will be considered frequently by others.

Among the points noted on leafing through the volume, the following may be worthy of mention. But it must be emphasized that any adverse criticisms are entirely trivial in comparison with the superb nature of the entire work. The figures are devoted to morphology; there are only two optic diagrams of the type which are such a valuable feature of Winchell (1951). The sign given for the extinction angle is the opposite of that called for by Schuster's rule; this must be remembered when comparing data with Winchell. On page 141 calcite is listed as the low-temperature stable polymorph; on page 187 this honor is awarded aragonite. Instead of using *Z* (as done by Berman, 1937, p. 354) to represent the tiny cation of co-ordination 4 (in silicates, phosphates, sulfates), it stands for an additional (Winchell) or *fremde* (Strunz, 1949) anion (see classes 30 and 31 on p. 393, or 41 and 42 on pp. 656–659). Kutnahorite is given on page 207 as $CaMn(Co_3)_2$, on page 213 as a dolomite with Mn:Mg = 1:1, and on page 217 as a dolomite with Mn > Mg or Fe (it is not listed in Winchell). In the figures of gypsum on page 483, *M* should be *f*. The andalusite formula on page 859 is wrong. Tungstates (pp. 1063 ff.) should now be called wolframates. Strunz (1949, pp. 181–184) has reoriented triphylite (p. 665) to agree with clinohumite (although he might better have chosen chondrodite); it seems unfortunate that the new Dana uses a new orientation (the old one appears in Winchell), apparently blindly following the $c < a < b$ rule.

The book would be improved if the mineral number appeared in the inside upper corner of each page. By adding tabs marking the 13 families (large type, Table of Contents), minerals lists are made much more accessible.

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The Tectonics of Middle North America. By PHILIP B. KING. Princeton: Princeton University Press, 1951. Pp. xix+203; figs. 52. $3.75.

This interestingly written book gives a condensed summary of the structural features of continental North America outside the Rocky Mountains and exclusive of the northern areas. It is an outgrowth of Dr. King's continued interest in this field which resulted in a very useful, but shorter, analysis of the structural geology of the United States, in connection with the sixteenth International Geological Congress, 1933. The author hopes to publish in due time a volume that will include the entire continent. The immediate reason for publishing the present volume, which is planographed and without any larger maps, was the author's desire to provide a text to accompany the several geological and structural maps of North America that have been published in recent years.

The southern portion of the Canadian Shield is treated very briefly, but the nature of the intricate structural and creation problems is brought out. Next are taken up the Central Lowlands with their plate of essentially flat-lying sedimentary rocks. The stratigraphy of the Paleozoic, Mesozoic, and Tertiary systems is sketched briefly, and the discussion of the various basins, domes, fault zones, and disturbed margins is accompanied by numerous structure sections. Seventy-eight pages deal with the Appalachian system of disturbed belts. It is gratifying that the author devotes as much space to a discussion of the structure of the crystalline rocks as to that of the fossiliferous sedimentary rocks. Areas that have been investigated in detail are portrayed, and the author has endeavored to interpret the structure of less-well-known districts by drawing cautious generalizations from the better-known ones. On the whole, the author's interpretation is commendably impartial, and alternative or diverging views are stated frankly.

The Ouachita-Marathon belt of folded Paleozoic rocks is discussed in 12 pages, followed by the concluding chapter of 22 pages on the structure of the Coastal Plains.

Each chapter has a number of well-chosen structure sections, many of which have been specially constructed to maintain a uniform scale. Outline maps are added for a few areas. At the end of each chapter is a list of references, about as up-to-date as can be expected and selected with great care.

The style is clear and the treatment so well balanced that one can hardly distinguish those areas in which Dr. King has done most of his own detailed work from others which he presumably knows from the literature only or from discussion with other workers. It is much to be hoped that the promised complete volume will be published soon. To students of structural geology, here and abroad, it should be as helpful and stimulating as is the present, preliminary volume.

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Emanuel Kayser's Abriss der Geologie, Vol. 1: *Allgemeine Geologie.* By ROLAND BRINKMANN. 7th ed. Stuttgart: Ferdinand Enke Verlag, 1950. Pp. viii+296; figs. 196.

Emanuel Kayser, for many years professor at the University of Marburg, Germany, published a four-volume textbook of geology and a shorter outline which have both gone through many editions during the twenties and thirties. After Kayser's death and under the difficult postwar conditions, Professor Brinkmann has undertaken to revise this work. The volume under review is an outline of physical geology, said to differ but little from a preceding edition which the reviewer has not seen.

Brinkmann has done an excellent job. All phases of physical geology are equally well presented. The text is, of necessity, brief, but numerous graphs, diagrams, tabulations, and excellent photographs give a remarkable amount of interesting information taken from the latest European literature available to the author. That many recent non-European references are lacking is understandable but hardly diminishes the value of the volume. Lists of references at the ends of chapters cover the literature of central Europe through the middle forties, and an Index of 23 pages is very helpful.

R. BALK

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Interpretation of Ore Textures. By EDSON S. BASTIN. ("Geol. Soc. America Mem.," No. 45.) New York, 1950. Pp. ix+101; figs. 28; pls. 18. $1.75.

This is the first work of American authorship to discuss in detail the subject of the interpretation of ore textures since the report of the Subcommittee on Mineral Paragenesis of the National Research Council, of which Dr. Bastin was chairman, was published in 1931. The book is highly welcome, because the author is noted for the excellence of his work on the interpretation of ore textures from a wide variety of metallic mineral deposits. It follows by a little over three years *Textures of the Ore Minerals* by A. B. Edwards, of Australia; both books treat much the same subjects but in different ways and with different degrees of emphasis on the various subjects. No student of ore deposits, no matter what his specific interests, can afford to be without either book.

Interpretation of Ore Textures begins with a short Introduction which is concerned mainly with the definition of terms to be used throughout the work. The first chapter considers the textures of magmatic ores, and Bastin emphasizes the fundamental requirement of such ores —"that their valuable minerals are direct crystallizations from magmas and are not later . . . replacements of solid igneous rocks accomplished by solutions from some external source." The unmixing of solid solutions is considered only briefly (3 pp.), in contrast with the 59 pages which Edwards devotes to the same subject.

Chapter 2 takes up the textures indicative of open-space deposition from true solutions, considering crustification, cockade or ring structure, fillings of minor fractures, pore fillings, and plant-cell fillings. The third chapter leaves the subject of ore textures to consider the principles of colloidal deposition as a preparation, needed by most economic geologists, for the discussion of colloidal textures which follows in chapter 4. The subjects of these two chapters are barely mentioned by Edwards.

Bastin, in chapter 4, suggests expanding the use of Rogers' term "colloform" to include not only those substances deposited as colloids in open space but also to include replacements carried out by particles of colloidal size. Granted that replacement can be effected by particles of this size, such an amplification of Rogers' definition is completely justified and actually would simplify any discussion of this phenomenon. There has been, however, no theoretically sound mechanism thus far put forward which satisfactorily explains how such a process can be accomplished. As the information presented in Bastin's book is primarily descriptive and observational rather than theoretical, it is not to be expected that he should consider the theoretical aspects of colloidal replacement, but the problem is, nevertheless, a very real one and, sooner or later, must be dealt with.

That there is transportation in, and deposition from, colloidal suspensions under natural conditions all geologists probably are willing to admit. There is, however, a very considerable question as to whether colloidal-sized particles ever did any replacing, in the strict sense of that term, at all. Replacement is a process of essentially simultaneous exchange of ions between the solid host and a fluid solution in contact with it, in which process the original volume of the host does not appreciably change. It is highly doubtful whether such a process can be accomplished by particles of colloidal size (even relatively water-free suspensoid particles), because they could not diffuse through the solid host and any already emplaced rim of guest material without completely disrupting the crystalline structure of both. It would seem reasonable, therefore, that the colloform textures (in Bastin's expanded use of Rogers' term) in ores formed by replacement must be due to other causes than the introduction into the host rock of colloidal particles of the guest. If this is true, then it must also be considered a possibility (despite the much greater theoretical likelihood of colloidal deposition by open-space precipitation than by replacement) that the colloform textures in ores deposited in open space may be caused by factors other than flocculation of colloidal suspensions. For, if colloform textures can be developed by ionic replacement reactions, it is also possible that such textures can be formed by ionic precipitation in open space, and one way in which they may be produced is probably suggested to anyone who has ever precipitated aluminum and ferric hydroxide in his study of elementary qualitative analysis.

Even less tenable than the concept of replacement by suspensoid particles is Lindgren's idea of gel replacement, which Bastin mentions on page 28. If limestone were to have been replaced by an equal volume of silica gel, as Lindgren suggested happened at Tintic, at least half the gel volume would have been water, which would have been driven out when the silica was crys-

tallized to quartz. This would have reduced the volumes of the replaced portions of the formations to half or less of what they were prior to the initiation of replacement, a phenomenon which is not observable in the Tintic district, where the replacement ores occupy essentially the same volumes as the limestone for which they were substituted.

As Bastin is careful to point out (p. 26), "little information is available concerning the water content of artificial flocculates from suspensoids"; and it may be that their water content is so low that the volume changes resulting from their transformation to the crystalline state are negligible. In many deposits formed by open-space filling, which show what are considered to be good colloform textures, these textures are exhibited by both sulfides and quartz. The sulfides, if they were transported in the colloidal state, almost certainly were carried as suspensoids and were flocculated as relatively water-free particles. On the other hand, it has not to my knowledge been suggested that the colloidal silica, from which such quartz is supposed to have been developed, was transported in any other manner than as an emulsoid or was precipitated in any other form than as a gel, in which the water content must have been high. Yet the colloform quartz associated with colloform sulfides shows no evidence of having undergone any greater volume change than the sulfides themselves. The so-called "syneresis cracks," where they exist, are appreciably no more abundant or sparse in the ore minerals than in the quartz gangue, an impossible condition if gel dehydration alone is the cause of these cracks. In many instances, in fact, there is little evidence of such cracks in either class of colloform mineral, a rather strong argument against gel deposition, if not against suspensoid flocculation.

It appears probable, then, that the presence of colloform textures alone is certainly not definite evidence that the material which exhibits these textures was transported in either of the two possible types of colloidal dispersion. It remains, of course, highly desirable to explain why colloform textures should develop in open space and replacement deposits. But this much seems certain: that at least some of such textures developed in open space may, and all formed by replacement must, owe their origin to one or a combination of several of the many variables which control the texture of minerals deposited by replacement or direct precipitation from hydrothermal fluids, and not to the flocculation of colloidal particles.

Bastin (p. 25) says: "While it is now generally accepted that certain ores were deposited as colloids the proof of such an origin in many cases offers inherent difficulties." And (p. 27): "By no means all of the rounded, more or less spheroidal forms found in ores indicate colloidal deposition." Such statements indicate that the author is not entirely happy with the concept of colloidal transport and deposition of ore and gangue minerals, and it seems to me that a complete re-examination and re-evaluation of the whole subject is definitely in order.

In chapter 5, Bastin takes up the subject of replacement textures, beginning with a definition of the process of replacement itself. His definition and those of Edwards and of Bateman in *The Formation of Mineral Deposits* (p. 127) are essentially the same, all three emphasizing the solution of a definite but small amount of the host and the essentially simultaneous deposition of the same volume of the guest mineral or minerals in its place. Any disagreement I may have with a definition accepted by so many geologists is certainly not a criticism of Bastin's book, which does not profess to be a treatise on the physical chemistry of replacement. Nevertheless, it seems to me that the idea that new open space is developed, if only momentarily, in the replacement process must be re-examined. The actual substitution of one mineral for another almost certainly cannot be the result of the immediate filling of open space developed by solution of a small amount of the host. If this were the mechanism involved, the formation of the first layer of the new mineral around the old would act as a definite barrier to further replacement. Rather, it seems probable that the replacement process is one of solid diffusion; the unwanted ions of the host diffuse out to the nearest contact of solid and fluid, and the new ions diffuse in from that surface at the rate necessary to maintain approximate electrical neutrality within the solid phase. Metasomatism must often require extensive changes in the structure of the host to convert it to that of the guest, but this process of structural change, in my opinion, is one which goes on entirely in the solid state. The physical role of the ore fluid is simply to bring to the solid-fluid interface the ions needed to insure the development of the guest and to carry away the discarded ions of the host. As an example, the replacement of covellite by chalcocite is not the filling by

freshly deposited chalcocite of tiny voids made in the covellite only the instant before but, instead, is the addition of half the number of cuprous ions and the removal of one-fourth of the sulfur already present in the replaced volume of covellite (see eq. [1]):

$$\underset{(136.96\text{Å}^3)}{4CuS + 2Cu^{+1} + 4Fe^{+3} + 4H_2O} = \underset{(136.62\text{Å}^3)}{3Cu_2S + SO_4^{-2} + 4Fe^{+2} + 8H^{+1}}, \quad (1)$$

Volume difference = 0.25 per cent.

After his short discussion of the process of replacement in general, the author takes up replacement textures under the headings of "Pseudomorphic," "Transecting," "Guided Penetration," "Diffuse Penetration," and "Automorphic." This section is followed by short discussions of selective and nonselective replacement, graphic textures, disparity in grain size between guest and host, form of guest-host contact, hypogene versus supergene replacement, and synthetic replacements. The discussion is complete and clear; it is to be regretted, however, that the author did not include a general summary treating the relative diagnostic value of the criteria proposed for replacement. Edwards approaches the problem of replacement textures in a somewhat different manner from Bastin, but the opinions expressed are much the same.

Chapter 5 considers microscopic inclusions to be found in ore minerals and the textures they exhibit. The author points out that very similar microscopic inclusions can form from a variety of processes and makes no claim for infallible criteria to distinguish one kind from another. So far as solid inclusions are concerned, Edwards covers this subject in part under "Solid Solution" and in part under "Replacement Textures."

Chapter 7 deals with the determination of age relations of ore and gangue minerals from the ore textures observed in polished section or, as McKinstry puts it, "[the] translation of textures into paragenesis." A discussion of the preparation of paragenetic diagrams also is included. The general subject of paragenesis is treated in somewhat more detail by Edwards.

Chapter 8 discusses the textures of sedimentary ore deposits, dealing mainly with the oölitic textures exhibited by so many iron ores. Chapter 9 considers those textures formed by dynamometamorphism, the author pointing out that, while "regional metamorphism is unfavorable for the formation of ore deposits . . . [it has] developed pronounced foliation in the ore." Chapter 9 is followed by a little over three pages of references, to which citations are made in the body of the text.

The book is completed by 18 plates, opposite each of which is a brief explanation of the textures depicted, and by a two- and a-half-page Index. The plates are nearly all well reproduced and well chosen except, perhaps, for too great a representation of silver-calcite ore textures.

Interpretation of Ore Textures will continue for many years to come to be of great value to students of ore textures and of ore deposits in general. To gain the broadest coverage of the subject, I would suggest its use in conjunction with *Textures of the Ore Minerals* and the section on "Mineragraphic Studies" in McKinstry's *Mining Geology*.

Dr. Bastin deserves the thanks of the profession of ore geologists for producing in his retirement such a concise and illuminating summation of his years of work with ore textures. The Geological Society of America is also to be congratulated for publishing a book which is of appeal to too limited an audience to justify its acceptance by a commercial publishing house.

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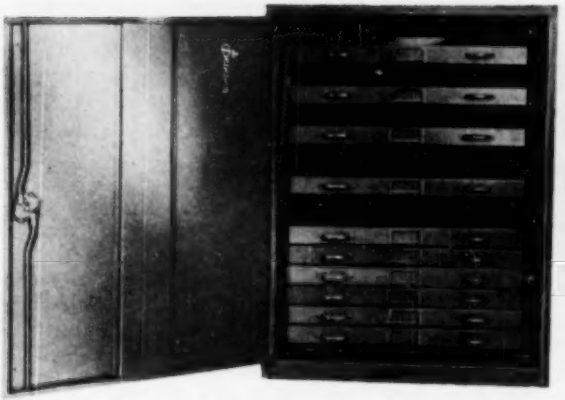

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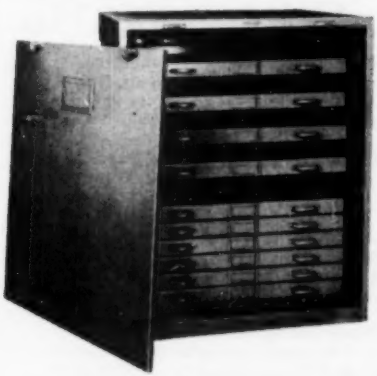

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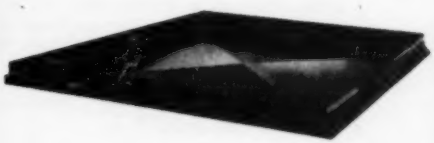

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